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REFERENCE

Lane, B. H., 1935, Suggestions to authors of papers submitted for publication by the United States Geological Survey: U. S. Geol. Survey, Washington, D. C. (For sale by Superintendent of Documents, Washington, D. C. 35c.)

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SEDIMENTATION PHENOMENA IN KULM AND FLOZLEERES GRAYWACKES, SAUERLAND AND OBERHARZ, GERMANY

Ph. H. KUENEN and J. E. SANDERS

ABSTRACT. Sedimentary structures and current-direction phenomena of the Kulm (Lower Carboniferous) and Flözleeres (Upper Carboniferous) graywackes of the eastern Rhenish Schiefergebirge (Sauerland) and Oberharz areas, central Germany, have been examined in reconnaissance. These rocks are interpreted as deposits made by turbidity currents in a relatively deep marine trough whose deepest portion migrated slowly northwestward across the geosynclinal tract. Current-direction phenomena support previous interpretations that a highland source existed during the Carboniferous in the present site of the Hesse graben. This highland transversely divided the elongate trough and provided coarse and fine sediment which was transported away on opposite sides, both to the northeast and southwest. Most of the turbidity currents traveled parallel to the length of the deep troughs, but a few crossed the Oberharz trough, apparently set off from the steep sides. Shallow-water deposits of the same composition as the graded graywackes occur along the southwestern side of the Hesse highland. Turbidity-current directions in the deeper-water areas further southwest are found to be directly away from these shallow-water deposits.

Features previously cited in favor of a littoral marine or deltaic origin for these graywackes: (1) coarse conglomerates, (2) ripple marks and ripple cross-lamination, (3) presence of plant fragments, (4) animal tracks and burrows, (5) preconsolidation distortion of bedding, (6) bottom-bed markings, (7) occurrence of radiolarian cherts, and (8) presence of fresh feldspar, are shown to be more consistent with a deeper water, turbidity-current origin, or irrelevant for environmental determination. The coarse conglomerates are thought to have been emplaced by submarine slides.

A previously unreported type of bottom-bed marking, designated *slide marks*, is described and compared with drag marks. Slide marks are formed along the basal plane of a slumped mass and might be likened to giant "slickensides" made during the movement of soft sediment masses one over another.

Graded bedding is present in many exposures, but is not common within a belt 10 to 15 kilometers wide parallel to the inferred shoreline. Graded beds occur up to 30 kilometers away from the outer edge of this non-graded belt. Assuming a 5° bottom slope for 10 kilometers, and 1° for the next 30 kilometers, a depth of 1500 meters is attained. This figure is believed to be a minimum depth estimate for the deeper parts of the trough.

INTRODUCTION

The interest of the senior author in sedimentary deposits characterized by graded bedding grew out of his studies of turbidity currents. He has examined numerous field occurrences of graded graywackes and other graded rocks and has enumerated the criteria by which the action of turbidity currents may be recognized and their paleogeographic significance assessed. Descriptions by Fiege (1936, 1937, 1951) of rhythmic graded bedding and other sedimentary features in the Flözleeres (barren lower part of the Upper Carboniferous) and Kulm[1] (Lower Carboniferous) of Sauerland and the

[1] "Kulm" as used in Germany refers to the flysch-like facies of the Lower Carboniferous. Both type of rock and geologic age are implicit in this usage. "Dinantian" is synchronous with "Kulm" but is applied to other rocks.

Harz, Germany, suggested the probability that these rocks, also, were deposited by turbidity currents. The authors therefore decided to visit the area to examine Fiege's localities and other exposures, in order to study the sedimentary textures of these rocks in an attempt to reconstruct their environment of deposition and to determine if current-direction studies could yield significant results (fig. 1). The Harz graywackes are in themselves a matter of interest because they are the original rocks to which the name "graywacke" was applied (Bailey, 1930).

From this reconnaissance study we conclude that these German graywackes can best be explained by the theory of turbidity-current deposition. It is the purpose of this paper to summarize previous theories of origin of these rocks, to describe the sedimentary textures and to interpret their meaning, to show the current directions ascertained, and to present the paleogeographic inferences made possible by these measurements.

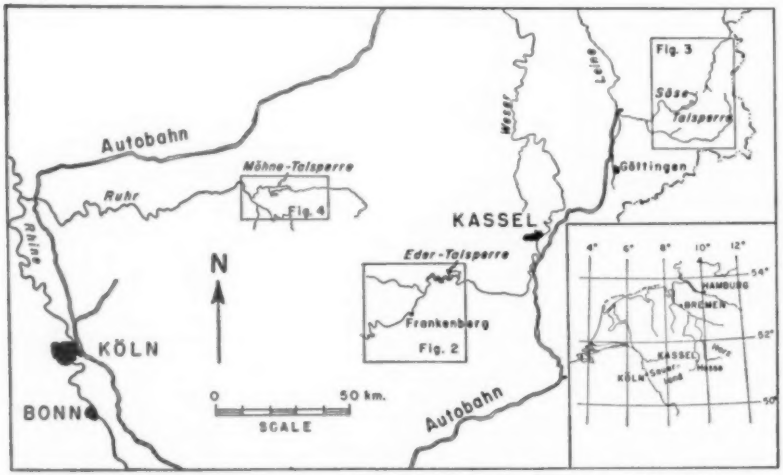

Fig. 1. Index map of west-central Germany showing location of detailed locality maps. Inset is sketch map of northwestern Europe to show regional setting.

ACKNOWLEDGMENTS

The authors express their thanks to Dr. F. J. Pettijohn for drawing their attention to Fiege's papers and to Professor C. W. Correns and Professor J. Tercier for assistance.

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PREVIOUS WORK

The general geologic setting of the graywackes of the Kulm and Flözleeres has been studied by numerous workers.

Schmidt (1923, 1924) established a standard goniatite sequence for the Carboniferous deposits of the Rheinische Schiefergebirge (of which Sauerland forms the eastern part) and Harz which serves as a framework for correlating and dating the deposits in widely scattered areas. Three zones, designated I, II, and III from base upward, and numerous subzones, denoted IIIα, IIIβ, IIIγ . . . IIIγ1, IIIγ2, etc., are recognized within the Lower Carboniferous.

Detailed stratigraphic and structural studies within the area of our reconnaissance have been made by Weigelt (1918), Hoeck (1929), Claus (1927), Stöcke (1929a, 1929b), Mempel (1933, 1935), Kobold (1933), Haubold (1933), and Pickel (1937). These authors have dealt primarily with stratigraphic subdivisions and correlation but some give consideration to sedimentary features and petrography. More general discussions of the regional relationships of the Kulm are found in papers by Born (1927), Kühne and Paeckelmann (1929), and Paul (1939, 1940). Locher (unpublished) has made a detailed petrographic and chemical study of the Kulm graywacke of the Sösetalsperre, Harz.

Fiege (1936, 1937) described the cyclic stratification and such sedimentary features as ripple marks, lower surface markings, contemporaneous deformation, etc., in the Flözleeres and Kulm, and plotted the map position of the graded and non-graded beds. He noted a repeated alternation of graywacke and shale beds and paid particular attention to the cyclic nature of these deposits. His work was done in the greatest detail and for several localities includes a list of all the beds exposed, their thickness, and particular properties. Fiege's inferences concerning the origin of these features will be discussed on a subsequent page.

GENERAL GEOLOGIC SETTING

The Kulm graywackes, of varying thickness, overlie finer-grained pelagic deposits of siliceous limestone, siliceous shale, and radiolarian chert (lydite). The passage from the fine-grained, pelagic deposits below to the massive graywackes above takes place through varying thicknesses of interbedded shale and thin, fine-grained graywacke beds. The oldest Kulm graywackes occur in the lower part of the uppermost subdivision of the Lower Carboniferous (Zone IIIα) and rise into successively younger subzones of Zone III (IIIβ, IIIγ) as one proceeds northward across the Schiefergebirge. Still farther northward and westward the Lower Carboniferous consists entirely of fine-grained deposits, and the graywackes occur in the Upper Carboniferous, dated variously as Lower and Middle Namurian and Upper Namurian (Lower Flözleeres) (Kühne and Paeckelmann, 1928). Some confusion exists in the older literature because the constant sequence of lithologic types from fine-grained beds below to coarse graywacke above gave rise to the conclusion that the similar rock units were coeval. The graywackes, for example, were all

thought to be equivalent to one another and were generally assigned to the Flözleeres.

Conglomerates occur at several horizons within the Kulm and Flözleeres graywackes.

The productive coal measures of the Ruhr basin are younger than the graywackes of the Flözleeres and occur along the northern edge of the Schiefergebirge.

These rocks were folded in the late Paleozoic (Variscan orogeny) and then beveled by erosion and unconformably overlain by flat-lying Permian and Triassic strata. Later northwest-trending normal faults (Hercynian movements)[2] cut the previously deposited rocks so that the Schiefergebirge and Harz are now separated by the Hesse graben, in which flat-lying Permian and Triassic are preserved.

PREVIOUS THEORIES ON CONDITIONS OF ORIGIN OF GRAYWACKES AND SOURCE OF SEDIMENT

Several authors have attempted to interpret the conditions under which these graywackes were deposited and to locate the source of the material of which they are composed. Most previous workers advocated a shallow-water origin and inferred some sort of pulsating supply to account for the repeated alternation of coarse and fine material. Various source areas have been suggested.

Conditions of Origin

Hoeck (1929) subdivided the Kulm of the Edertalsperre and assigned positions in Schmidt's zonal scheme to the radiolarian cherts, graywackes, and conglomerates. He also discussed the origin of these rocks. With reference to the radiolarian siliceous rocks, he remarked (p. 23) that these had been interpreted previously as deep-sea deposits (probably after the influence of Haeckel's classic work on the radiolarians collected by the *Challenger* expedition) but that others (Dixon, 1911; Davis, 1918; Correns, 1924) had later suggested a relatively shallow environment for the formation of these rocks. Hoeck (1929, p. 29) cites Wedekind as suggesting a depth of from 50 to 200 meters for the rocks containing goniatites and *Posidonomya*. Hoeck concluded that radiolarites as well as graywackes and conglomerates were deposited in shallow water near a coast. Evidence cited in support of this conclusion includes: coarse conglomerates, abundance of plant fragments, abundant fresh feldspar, ripple marks, and cross-bedding. These rocks were inferred to have been deposited on a delta, and the conglomerates to represent channel deposits.

Kühne (in Kühne and Paeckelmann, 1929) agreed with Hoeck on the nearness of the source area to the site of deposition and claimed that the rhythmic interbedding of coarse and fine material resulted from fluctuating intensity of supply, which might have been controlled by climatic variations or by periodic renewal and decay of the source area by intermittent uplift. The gradual northward transgression of the graywackes in time was taken

[2] This usage of "Variscan" and "Hercynian" is that of the German authors cited in the bibliography and differs from the usage originated by Bertrand which considers "Hercynian" as a Late Paleozoic orogeny. For further discussion, see Bailey, 1935, p. 5, 8, 13-14.

as evidence that the coastline shifted northward. Kühne and Paeckelmann also remarked on the similarity of the Kulm to the Alpine flysch.

Stöcke (1929a, 1929b) studied the Kulm conglomerate between Frankenberg and the Edertalsperre and also concluded that deposition took place on a delta. The interbedding of coarse and fine material was attributed to rhythmic movements of the source area (1929a, p. 1032).

Fiege (1936, 1937) described the cyclic deposition of the Kulm rocks over wide areas in graded graywacke beds alternating with shales. He laid much emphasis on the variable thickness of the beds and on the occurrence of longer cycles in which a number of beds show a gradual change in character from mainly clay beds becoming more and more sandy, until after a few coarse graywackes a new major rhythm sets in above with clay beds. For many localities he even published a list of all the beds with their individual properties. The reason for taking so many pains over the minutiae of the bedding was that Fiege attributed a deeper significance to them. He believed that for each graywacke bed an influx of coarse sediment into the geosynclinal trough took place and that the upward grading to the fine-grained top of the graywacke bed testified to the diminishing supply of coarse material. In a later paper (1951) he sought the cause in a climatic influence. The major cycles he attributed to epeirogenic uplift, each rejuvenation of the source area causing a series of coarser beds.

Fiege was able to confirm and extend the picture, already drawn by others, of zones parallel to the former coastline: starting in the east or south with shallow-water deposits, without grading and with coarser and more irregular bedding, passing outward into deeper-water beds consisting of graded graywackes and finally reaching the fine-grained Kulmtonschiefer (clays) and Posidonian Schiefer and siliceous shales of pelagic nature.

Mempel (1933) showed that in the Oberharz the Kulm starts with siliceous shale, is overlain by shale and this succeeded by graywackes, in which there occur three conglomerate zones. During the time that 50 meters of Posidonian Schiefer were deposited in the northwest, more than 400 meters of graywackes accumulated in the center in the Sösetal area. Total thickness of the Kulm is estimated at 1500 meters. Mempel assumed that the pebbles were carried to the coast in mudflows and were deposited in a delta on the sea floor. He estimates these conglomerate fans to have a radius of 150 km. They are found in Kulm IIIβ3, IIIγ1, IIIγ1/2, each successively younger one occurring further north.

Pickel (1937) restudied the Kulm of the Edertalsperre and attempted further analysis of the conditions of deposition. He decided that these rocks had been deposited in a foredeep and that the source lay to the south. He thought the graywackes were deposited in shallow water on a delta after the material had been transported down from the highlands by mudflows and that the great thickness indicated rapid subsidence (p. 264). The presence of ripple marks, worm tracks and tubes, and traces of subaquatic sliding were cited in support of the shallow-water hypothesis, and the abundance of plants adduced in support of the nearness of the coast. In addition, Pickel (1937, p. 271) observed and photographed the parallel markings on the

bottoms of certain graywacke beds and held these, also, to be indicative of deposition in very shallow water. He thought they were formed in rocks that were enough consolidated to be able to retain these markings when the strong currents brought in the coarse material which overlies them. He also suggested that these markings might be somehow related to an unspecified flowage effect during the drying-out of the overlying bed, but that drying could not have continued for long, as no mud cracks were observed.

Pickel further observed that if the "cycles" include graded bedding, then the upper part of the bed consists of micaceous or plant-rich shales; if non-cyclic deposition occurred, then the upper surfaces of the beds are ripple-marked; but he was unable to determine a definite ripple direction. Pickel's studies of the "cycles" in the Kulm resemble the studies made by Fiege in the Flözleeres, and similar conclusions were reached. The Kulm graywackes were found to contain more matrix and less quartz than the Flözleeres graywackes and the range of pebble sizes in both to be the same.

Pickel noted the graded bedding and sharp basal contact of the graywacke beds. He believed this to mean rapid pouring-in of weathered debris at the beginning of a "cycle" and then gradual diminution of supply as the cycle continued. At the end of the cycle the supply ceased. Climatic or tectonic causes produced this result. The distribution of the cyclic and acyclic beds was thought to indicate the position of the coast, as evidence of the cycles would be destroyed by reworking in the shore zone or on a delta but could be preserved a short distance from the coast in somewhat quieter water.

Source of Sediment

According to Born (1927) the pebbles of the Upper Carboniferous conglomerates in the Rhenish geosyncline are perhaps second- or even third-cycle fragments and might have been derived from the Kulm conglomerates. Ultimately, he argued, these pebbles in the Kulm conglomerates could not have come from the south but from a crystalline area in the "Variscan basement" to the north.

Kühn and Paeckelmann (1929) disagreed with Born and insisted the pebbles could not have come from the north, because in those areas (in Westphalia) the Kulm is fine grained and lacking in pebbles. The source was the area between the Harz and Schiefergebirge. They also pointed out that in Westphalia the oldest conglomerate occurs in the upper Middle Carboniferous, not in the Kulm, and that here the conglomerates of the Flözleeres contain pebbles of the Kulm lydite, indicating that the Kulm had been uplifted and subjected to erosion prior to deposition of the Flözleeres.

Stöcke (1929a) made a detailed petrographic study of the Kulm conglomerates of the Frankenberg-Edertalsperre area and found granite, quartzite, vein quartz, keratophyre, gneiss, porphyry, and other pebbles and cobbles. The decrease in grain size toward the southwest was taken as proof that the source lay to the northeast, under the Zechstein (Permian) and Mesozoic rocks of the Hesse graben, where he showed a Lower Carboniferous highland (p. 1032, fig. 2) forming a connection between the "North Atlantic Continent" and the "German Island." Because of pebble distribution in the conglomerates,

he thought the highland must have been composed of granitic rocks in its northern part and metamorphic rocks in the southern part. This same highland was regarded as the source of the pebbles in the Kulm conglomerates of the Harz. As additional support in favor of a northeastern source Stöcke also cited gravimetric measurements which suggest the presence of crystalline basement at shallow depths underneath the flat-lying Mesozoic and Permian rocks of the Hesse graben.

Fiege quoted Dahlgrün as claiming a supply in the Harz from the south and Mempel as from the south-southwest and agreed with the latter conclusion. The coastal area was apparently roughly in the neighborhood of Sachsa. Mempel (1933, 1935) opined that the Harz pebbles and those of the Kulm exposed 100 km to the northeast all came from a source area in the south-southwest.

Pickel (1937) studied the Kulm conglomerates in great detail and concluded from the northwestward decrease of thickness, maximum pebble size, average pebble size, percentage of pebbles, and percentage of crystalline rocks that the source generally lay to the southeast, both for the Edertal and Harz conglomerates. Pickel also claimed that the presence of limestone pebbles indicates a southern source. According to his interpretation, epeirogenic uplift of the source area pushed the shoreline northward, and the Kulm foredeep migrated to the north in time. He disagreed with Stöcke concerning the significance of the pebbles of igneous and metamorphic rocks. Pickel (1937, p. 264) indicates that the stratigraphically lower conglomerates are richer in metamorphic rocks and that the younger conglomerates contain more igneous material. Therefore, he argued, a plutonic body was being unroofed and few igneous rocks were available until deep erosion had laid bare the core.

Fiege studied the positions of the graded and non-graded beds in the Kulm graywackes in attempting to locate the source and coastal areas. He showed (1937, fig. 20) the non-graded graywackes to lie east of a line from the Edertalsperre to Kölzhausen, a direction 15° east of north. Fiege inferred that the supply in the Flözleeres came from the south-southwest.

According to Paul (1939) the Kulm represents deposits made in a great central European basin around the margins of which deposits of shallower water (Carboniferous limestone) were laid down. The coarse material of the Kulm came from islands within the basin.

PRESENT INVESTIGATION

Favorable exposures found in road cuttings and quarries were visited in a field reconnaissance made between 29 April and 5 May, 1954. Oriented sedimentation phenomena were studied and measured, and the presence or absence of graded bedding determined. The Flözleeres and Kulm were treated as separate, undivided units in the field. Although certain large sub-units appeared obvious, it was only after later search of the literature by the junior author that the established stratigraphic subdivision of these beds was discovered. To position each outcrop in the sequence would require a separate investigation.

Measurement of oriented phenomena was made by placing a rectangular notebook on the bedding plane concerned, with the long side of the notebook parallel to the strike. A pencil was aligned by eye parallel to the feature to be measured and made fast to the notebook in this position by elastic bands. The notebook was then rotated to horizontal about the edge held parallel to the strike. Special care was taken to determine the stratigraphic top of the beds in order to rotate the notebook by the correct edge. With the notebook horizontal, a compass bearing was taken on the pencil and recorded to the nearest 5°. Where the oriented phenomena are parallel to the strike of the beds, as was found in many localities, the foregoing procedure is unnecessary and the compass bearing may be read directly.

Rotation back to horizontality around the tectonic strike is based on the assumption that simple rotation around the strike is the only pattern of deformation. Considerable departures from this assumed condition are to be expected in regions of intense tectonic disturbance. As flexure folding about gently plunging fold axes seems to be the chief style of deformation in the rocks examined by us, and criteria are lacking to determine the extent of any departure from the assumed conditions, we employed the procedure outlined.[3]

In reading ripple lamination care must be exercised because generally only oblique components of the true current direction are exposed on a single surface. Small pieces of the rippled bed were broken from the outcrops in order to inspect the inclined lamination in three dimensions. As a rule, ripple lamination is found easily in thin, slightly irregular "wavy" beds, and in some instances may be absent from other parts of the section.

Results of Observations

Kulm near Frankenberg and the Edertalsperre.—The Kulm of Sauerland was examined in 31 localities in the neighborhood of Biedenkopf, Hallenberg, Frankenburg, Frankenau, and the Eder-Talsperre (fig. 2). Most of these localities are outcrops of graywacke, but two are in the Kulm Kieselschiefer. These graywackes are the oldest examined and have been placed in Zone IIIβ.

Many sedimentation phenomena were observed in these localities. Graded bedding is very common. In nearly all exposures clear evidence of current orientation exists in the form of flow marks or drag marks, aligned plant fragments, etc. Except for two or three cases, ripple lamination or shape of flow marks showed which way the current had followed along these markings.

The siliceous shales present partings here and there which somewhat resemble ripple mark. No internal cross-lamination can be detected, however, not even on polished surfaces. Hence these partings are regarded as a secondary jointing phenomenon.

At locality 14, a small roadside quarry southeast of Braunshausen, the writers observed interesting phenomena, some for the first time within their experience. On the upper surfaces of non-ripple-marked beds, plant fragments occur with their long axes parallel to the current direction, as noted in most other localities.

[3] An ingenious device for restoring the original position of sedimentation phenomena in plunging folds has been designed by H. Cloos (1938, p. 360-364). Stereographic projections could also be used.

PLATE 1

A. Aligned coaly plant fragments on loose block. Kulm, Grosse Steimkertal, Harz (locality 48).

B. Aligned single and double pelecypod shells on bedding plane. Kulm, Frohnhausen, Sauerland (locality 18).

Locality 5, in the Flözleeres, at Niederense (see below), had provided our first encounter with aligned plant fragments (now carbonized) on the upper surfaces of the beds. Such carbonized plant fragments occur in great abundance towards the upper surfaces of some beds, as noted by Fiege, and were found by us to be oriented with their long axis more or less parallel to the direction of current flow ascertained from the other phenomena[4] (pl. 1-A).

On one especially well exposed bed of locality 14 containing large current ripple marks on its upper surface, plant fragments were concentrated in the troughs between the ripples with the long axes of the fragments aligned at 90° to the current direction. A few small fragments of plant material remaining on the crests of the ripples were lying with long axes parallel to the current direction.

Evidence of submarine slumping is also preserved in this quarry. The plane of contact between the slumped mass and stationary substratum just

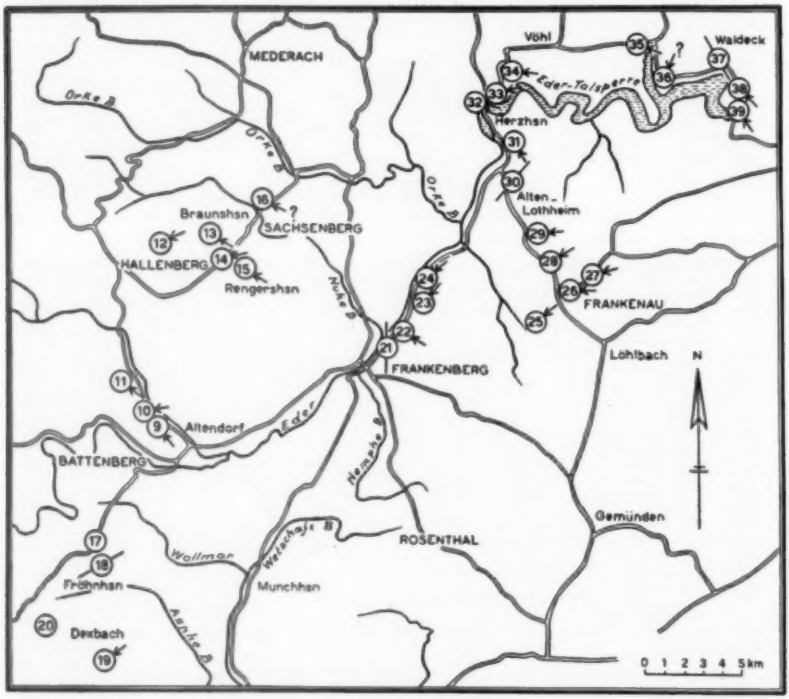


Fig. 2. Locality map of area near Frankenberg and Eder-Talsperre, Sauerland. Circled numbers designate localities. Arrows by each circle indicate direction of inferred current flow. Lines signify current path where no direction of movement could be proved.

[4] Oriented plant fragments have been used to infer current direction by H. Cloos (1938) and others. Crowell (1955, p. 1357, 1361) employed this feature in turbidity-current deposits independently before us in a study of the Prealpine flysch of western Switzerland made in the fall of 1953.

upslope from the distorted slumped mass reveals grooved markings, which the authors here propose to call "slide marks." They differ from drag marks in being absolutely parallel to one another, but not perfectly straight, and in covering the whole basal plane of the slumped mass (pl. 2). Drag marks commonly diverge somewhat from parallelism and are some distance apart or single. Until the writers discovered this clear proof of the origin of slide marks, they had interpreted similar markings in the Flözleeres as drag marks (specimen collected from locality 7).

Aligned pelecypod shells were discovered in fine-grained graywackes at locality 18 (Frohnhausen). The long axes of the single or double shells were parallel to the current direction (pl. 1-B).

Current flow read from the graywackes of this area is consistently from the east, though considerable divergence is found, ranging from 30° to 150°E (fig. 2). This result bears out the conclusions of Fiege (1937).

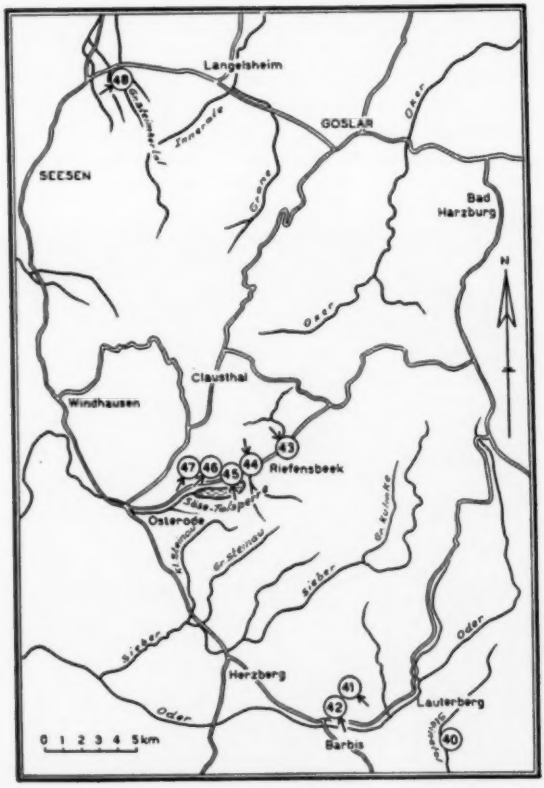


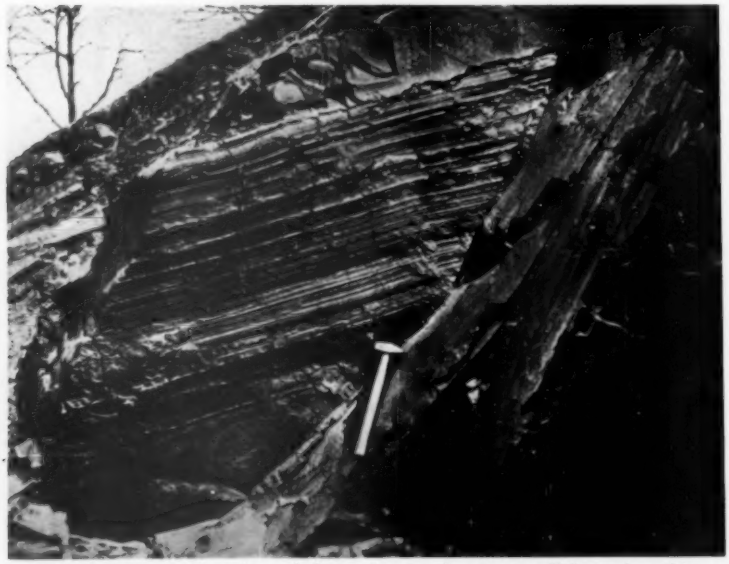
Fig. 3. Locality map of area near Söse-Talsperre, Oberharz. Circled numbers designate localities. Arrows indicate inferred direction of current flow.

At the Eder dam (locality 39) the writers observed coarsely cross-bedded, non-graded sandstones, which are irregularly bedded and contain scour-and-fill channel deposits. The material of these beds is similar in composition to the graywackes, but the sedimentary textures are strikingly different and clearly indicate deposition in shallow water. A coarse conglomerate containing well rounded pebbles and cobbles of granitic and other crystalline rocks, which we infer to be a slide conglomerate, occurs a few kilometers northwest of these shallow-water beds in a roadside quarry (locality 37). Grading is present in the upper beds of this quarry. These rocks are evidently foreset deposits or slumped foresets formed on the outbuilding slope of a delta advancing westward into deep water.

The presence of shallow-water beds in the direction from which the turbidity currents came is highly significant and represents the first case in which turbidity-current direction has been determined over a wide area which also includes the shallow-water deposits from whence the turbidity currents evidently originated. Carozzi (Kuenen and Carozzi, 1953) has described two cases in the Alps where graded beds have been traced to their origin, but without ascertaining flow directions.

Kulm of Oberharz.—The graywackes of the Oberharz were examined in 9 localities, 6 of which are of Kulm age and 3 (localities 40-42) belong to the Tanner graywackes of pre-Kulm age (fig. 3). The precise age of the Kulm rocks visited is not known to the authors, except for those in the Grosse

PLATE 2



Slide marks on basal surface of slumped mass, showing parallelism of markings and slight curvature of their course. Kulm, SE Braunshausen, Sauerland (locality 14).

PLATE 3

A. Polished surface showing entire thickness of fine-grained graywacke bed. At bottom cross-lamination, current from left; remainder with convolute bedding. Anticlinal shapes "heel over" in down-current direction. Kulm, Söse-Talsperre, Oberharz (locality 45).

B. Groove casts of drag marks and later casts of flow marks made by turbidity current moving from left to right. Slight faulting offsets markings. Base of graywacke bed. Kulm, Grosse Steimkertal, Oberharz (locality 48).

Steimkertal (locality 48) which were assigned to the "Upper graywacke" of Zone IIIγ by Kobold (1933). Mempel (1933) found conglomerates with ages IIIβ3, IIIγ1, and IIIγ1 to 2. If these ages are correct, then these graywackes are somewhat younger than those of the Frankenberg-Edertal region. In some of the Oberharz localities, studies for determining current directions are particularly rewarding and show, in addition, numerous other phenomena which are characteristic of turbidity-current deposits when found together in the same rock sequence, such as convolute bedding, graded bedding, drag marks, and flow marks. At locality 48 (Grosse Steimkertal, Oberharz) both groove casts (of drag marks) and later flow marks are present on the same bedding plane. The drag marks formed first and were later partially obliterated by flow marks (pl. 3-B).

Leaving the Tanner graywackes out of consideration, the dominant current direction is from the SSW (210°E, average loc. 45-48). However, at two localities (43 and 44) a NW direction (average 320°E) was clearly discerned. Locality 45 with a SSE reading may even represent a third direction of supply (fig. 3).

The clustering of three current trends about directions at right angles to each other is somewhat unusual, and several possible explanations could be applied. If the exact age relations of the beds concerned were known, they might substantiate the idea that tectonic activity had altered the bottom slope with time. In dealing with two directions of sedimentation approximately at right angles in the Prealpine flysch of Switzerland, Crowell (1955, p. 1370) discusses the concept that submarine canyons acted as point sources which localized the turbidity currents. Thus current directions would follow the radii of a submarine fan or delta in moving away from the mouth of a canyon. Some currents might by this means be directed parallel to the coastline. Though this concept is useful, the writers suggest that fans should cause current directions to be approximately equally scattered within a semicircular area, and not direct them in one or more preferred directions. Many observations, of course, would be required to test this hypothesis. A third alternative, preferred by the writers, is that the Oberharz area examined lay on the nearly flat floor of a steep-sided, U-shaped trough; that turbidity currents flowed from the closed part of the "U" and also down the steep sides. Such morphological control by three mutually perpendicular downslope directions best seems to explain the three trends, though it must be admitted that further observation might show these apparent trends to be unreal statistically. Under favorable circumstances detailed provenance studies might show whether one source or several different ones contributed to the sediment.

Similar diverse current directions might be expected in a modern trough such as the Adriatic Sea, where the Po delta provides material for possible turbidity-current distribution parallel to the length of the trough, and where the steep lateral slopes off the Italian and Yugoslav coasts might periodically send down smaller currents which could locally produce records of current movement across the trough.

Consideration of the regional relationships of the Kulm of the Oberharz and the northeast end of the Rheinische Schiefergebirge indicates the Hesse

graben, which preserves Permian and Triassic flat-lying rocks and separates the two areas of Kulm outcrop, was a landmass and a locus of shallow-water deposition during part of the Lower Carboniferous. Pebbles in the Kulm conglomerates are thought by Stöcke (1929a) and others to have come from this tract. Current directions observed by us, both near Frankenberg and in the Oberharz, on opposite sides of this positive area point directly away from the shallow-water deposits and thus support Stöcke's conclusion.

The writers therefore suggest that two deep troughs separated by land existed on the sea floor in the Lower Carboniferous. The biostratigraphic work, however, indicates that the graywackes deposited in these troughs are not precisely correlative, so the troughs may not have coexisted in time.

Flözleeres near the Möhnetalsperre.—The graywackes of the Flözleeres were examined in 8 localities east and west of the Möhne reservoir (fig. 4). In general, these rocks consist of fine-grained graywacke in thin to thick beds, interbedded with shale. Oriented current phenomena are not abundant here, but current directions were read from the ripple lamination in 6 localities. Drag marks preserved as groove casts on the bottoms of the beds were also found at these 6 localities. Other bottom markings were interpreted as flow markings but are not distinct enough for reliable readings of current direction.

Our brief reconnaissance of the Flözleeres indicates nearly consistent current flow from an eastward direction (fig. 4). One exception occurs west of Völlinghausen (locality 7) where ripple lamination in three or four beds indicates current movement from the west-southwest, nearly opposite to the other observations. Here again a reversal of bottom slope or flow in two directions on an almost horizontal trough floor might explain the anomaly. The latter assumption would require two similar source areas, however, as the appearance of the Flözleeres at all localities examined is much alike, though the rocks have not been examined microscopically.

Present Interpretation of the Graywackes

Certain aspects of previous interpretations of the Kulm and Flözleeres graywackes must be abandoned in the light of more recent ideas on the nature

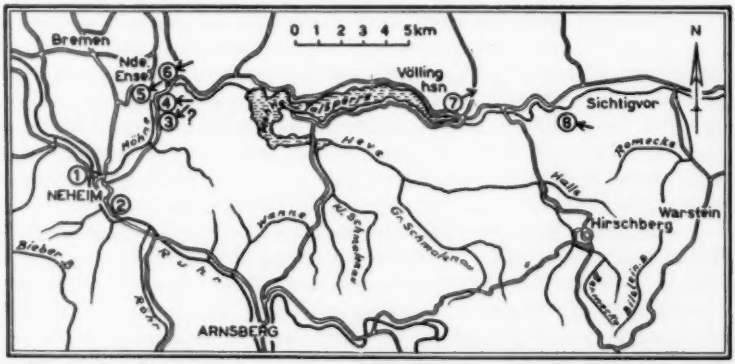


Fig. 4. Locality map of area near Möhne-Talsperre, Sauerland. Circled numbers designate localities. Arrows beside circles indicate inferred direction of current flow.

of deposits formed by turbidity currents, as has been shown in several publications by the senior author, partly in collaboration with Migliorini and others (see References).

The association of abruptly alternating coarse- and fine-grained beds through great thicknesses of very regularly bedded strata in the Kulm and Flözleeres, the graded arrangement of the material within many of the graywacke beds, the small variation in direction of current flow, indicated by ripples and various markings on the soles of the graywackes, together with the absence of coarse cross-bedding, represent a combination of physical features which is unexplained by previous interpretations. The theory of pulsating supply governed by processes operating in a source area on land is unsatisfactory, for, as pointed out by Fiege, material brought to the sea by rivers first encounters shallow water, where waves and currents constantly rework it. Any sediment which reaches deeper, calmer water must have passed this intervening zone of reworking, in which all evidence of any short-period pusation of the supply from the land source would be obliterated. At most, changes in the average thickness and grain size of the coarse beds may record alternating periods of rejuvenation of relief and its being subdued. It is also obvious that rejuvenation of supply would have to be instantaneously at its maximum to form a graded bed, because the stratum starts with the very coarsest grain. The dwindling, on the other hand, would have to be a gradual process. And yet this gradual process is entirely free of fluctuations, such as would be induced by tidal or wind-driven currents, by variations in weather or climate. Manifestly the emplacement is too rapid to be influenced by such factors, and also for benthonic life to survive, because only in the shales are traces of life in situ preserved.

The many arguments in favor of turbidity currents cannot be repeated here in detail. But the following points may be recalled. Indications of some kind of current action are ubiquitous in the form of flow- and drag marks below the coarse beds, current ripple lamination and oriented plant remains at the top, and oriented grains throughout. These currents tend to show a constant direction. The absence of wave ripple mark, of coarse cross-bedding, and of channel scour, the frquency of slump structures and of shallow-water organisms in reworked condition only, and the scarcity of benthonic life other than worms in the shales together indicate a deep, poorly ventilated environment. The load casting and convolute bedding signify extremely rapid deposition. Grading of the coarse fraction and alternation of coarse and fine sediment throughout the whole thickness is typical of deposits from experimental turbidity currents and of natural graywackes alike.

The pulsating supply indicated by the details of the stratification of the graywackes is attributed in the theory of turbidity currents to periodic development of slides and turbidity currents on steep, embankment-type[5] submarine depositional slopes. The concept of deltaic deposition advocated by previous workers may be accurate in a broad way but requires modification, as will be further shown below.

[5] An apt term proposed by J. E. Marr (1929).

All features of the Kulm and Flözleeres cited by previous workers in support of shallow-water origin can be explained at least as satisfactorily if not better by deposition from turbidity currents below wave base. A list of these features, with discussion of the value of each, follows:

The coarse conglomerates.—The coarse conglomerates interbedded with the graywackes have been taken by all previous workers as reliable criteria of shallow marine waters or stream channels. The properties of these conglomerates and the petrographic composition of the pebbles and cobbles have been adequately described previously and need not be repeated here. The rounding of the fragments indicates transport by torrential rivers or rolling on a beach prior to final deposition.

Many very coarse, non-graded conglomerates tend to look alike and may reveal little about their origin in isolated exposures. Considered in relation to adjacent strata, however, significant inferences may be possible. In the cases we encountered, the conglomerates are interbedded with graded graywackes, which we claim were deposited by turbidity currents at greater depths than those at which ordinary waves and currents are capable of moving coarse debris. No cross-bedding occurs in the conglomerates and some beds even pass upwards into a graded graywacke at their very top. We therefore infer that the conglomerates were emplaced by slides arriving in the deep-water environment. The boulders were presumably brought to the sea by swift mountain streams from a steep mountain chain, perhaps onto a delta built into deep water. Some of the beds containing coarse debris may be steep foresets. Others may have resulted from slides passing out onto the less steep sea floor beyond. At any rate the presence of graded bedding argues against their being channel deposits on the topset portion of a delta, or beach deposits. Therefore, we challenge the validity of coarse debris as a criterion of shallow-water deposits.

Current ripple marks and small-scale current ripple lamination.—"Ripple marks" and "cross-bedding" have also been regarded previously as reliable indicators of "shallow-water" deposits. The validity of such generalized usage must be questioned, however, and distinction made between the various types of ripple marks and cross-stratification (McKee and Weir, 1953). In shallow water both current ripples and wave ripples form. The latter appear to be restricted to shallow depths; the former may occur in water of any depth provided there are currents which can move the bottom sediment. Current ripple mark has been observed by the senior author in hundreds of beds in graded series the world over which preserve textures that cannot have been disturbed by ordinary shallow-water currents. Not a single unequivocal case of wave ripple mark has been found after careful search of thousands of beds in these graded deposits. This applies equally to the Kulm graded sequences examined, and it can be safely claimed that all ripples reported by others from graded portions of the Kulm are of the current type. These ripples all occur in the finer parts of graded beds and prove that currents deposited these beds. The general uniformity of current direction and associated features point to turbidity currents as the depositional agent. Obviously, then, current ripple marks and small-scale ripple cross-lamination are not unique

criteria of shallow depth; on the contrary they may be expected as well in deposits laid down by turbidity currents in deep water.

The fact that turbidity currents cause traction (ripple marks, cross-lamination) on the bottom over which they pass urges extreme caution in classifying sediments into two groups on the basis of traction-type vs. turbidity-type of depositing currents, as suggested by Packham (1954). The authors agree that distinction can generally be made between the two contrasting environments, as noted earlier by Bailey (1930), but the term "traction" for one group seems ill chosen because of the obvious traction effects also associated with turbidity currents.

The long axis of current ripples formed by turbidity currents is on the average at right angles to the current-flow direction, but the sinuous trend of the crests may result in local divergencies of 45° or even more. In one case (locality 19) we found the long axis of indistinct ripples parallel to the current direction. The internal lamination of these ripples indicates sand transport parallel to the long axis of the ripples. The conditions under which such ripples form is unknown to the authors. Perhaps some analogy could be made between these ripples and longitudinal sand dunes. The longitudinal ripples discovered by van Straaten (1951) on tidal flats are quite different, being features of erosion only, rather than products of erosion and deposition. Moreover, they result from the combination of wind waves playing on a current in shallow water.

Presence of plant fragments.—In the present study fragments ranging in size from leaves and twigs to tree trunks were encountered, locally in great profusion. Remains of plants have been taken almost universally by previous workers as an indication of non-marine or perhaps littoral deposition. Large rivers carry vast quantities of drift wood, leaves and stems out to sea, however, as Lyell mentioned more than 100 years ago (1830, p. 189). At sea these eventually become waterlogged and sink to the bottom. This may happen directly in deep water or in shallow coastal waters. In the latter case the fragments could be transported in a turbidity current and redeposited in deep water. Deep-sea expeditions have encountered pieces of wood on the present deep ocean bed, and Natland deduced a depth of 300 to 800 meters from foraminifera for beds containing charcoal in the Pliocene of the Ventura Basin (Natland and Kuenen, 1951).

Animal tracks and burrows.—These phenomena are not limited by depth. Submarine photography has shown that they are developed on the deep-sea floor, and they may be readily observed on tidal flats.

The frequency of worm tracks on the shale beds where all other indications of benthonic life are absent argues in favor of very low oxygen content of the bottom waters, and that in turn means stagnation between turbidity currents.

Loadcasting, convolute bedding, and pre-consolidation slumping phenomena.—Distortion and deformation of bedding are not criteria for depth of water or mechanism of deposition, but rather are related to the physical properties of the coarse and fine sediment and to depositional slopes, respec-

tively. These phenomena can be used under favorable circumstances to infer direction of depositional slope (pl. 3-A).

Markings on bottoms of beds.—Parallel and subparallel flow markings and drag markings (pl. 3-B) on the bottoms of beds observed in the Kulm and Flözleeres are identical with those of other graded series and are produced time and again by turbidity currents in deeper water, as indicated previously by Rich (1950), and the senior author (1953a, b). They do not form in the intertidal zone as suggested by Pickel (1937).

Siliceous shales containing radiolarians.—Rocks of this type were cited long ago as proof of deposition in abyssal ocean depths, but later it was thought they could have formed in shallow seas. Association with coarse-grained rocks and presence of ripple marks are the most important arguments which were previously cited in favor of a shallow-water origin. As we have indicated on a foregoing page, however, these features by themselves are no longer reliable criteria of a shallow-water environment; the present interpretation of the Kulm graywackes reopens the possibility of deep-water origin for the underlying siliceous shales, although no precise depth limits can be assigned.

Presence of fresh feldspar.—Fresh feldspar in a sedimentary deposit indicates only that mechanical disintegration and rapid transport have reduced the relative importance of chemical decay in the source area and cannot be utilized to indicate depth of water at the site of deposition.

Paleogeography and Geologic History

Our reconstruction of the paleogeography and geologic history of the Sauerland and Harz areas is as follows.

During the Lower Carboniferous a relatively deep trough existed in central Germany. At first no coarse sediment arrived in the area now open to investigation, and only pelagic rocks (chert and fine-grained shale) were deposited. The Harz and Sauerland which are now separated may have been part of a continuous trough during deposition of these pelagic rocks.

Near the end of the Lower Carboniferous, a highland of continental (metamorphic and granitic) rocks appeared in the trough between the Harz and Schiefergebirge, where the Hesse graben now stands. Great quantities of clastic debris were shed down both sides of this landmass, which transversely divided the basin. Turbidity currents were set off on steep submarine slopes and carried coarse material far out along the more gently sloping floor of each trough, passing generally from the northeast in the Schiefergebirge, and from the southwest in the Harz (though locally material was transported down the sides of the Harz trough from the northwest and southeast). These currents deposited the graded graywackes. Farther southwest in the Schiefergebirge (in Westphalia), pelagic conditions were undisturbed by the turbidity currents from the northeast, and siliceous shale and fine-grained shale continued to accumulate undiluted by coarser grades. The Hesse highlands supplied coarse boulders to the shore and these were caught up in slides down steep underwater slopes to become interbedded with the graded graywackes.

Once graywackes began to accumulate, the deepest part of the trough migrated slowly northwestward, probably partly by infilling, partly in response to compressional forces which tended to fold the Schiefergebirge, causing successively younger graywackes toward the northwest. In Westphalia a deep trough persisted well into the Upper Carboniferous in which the Flözleeres graywackes were deposited, apparently derived from the same northeastern source as the Kulm graywackes. Pebbles of Kulm lydite in the Flözleeres conglomerates indicate tectonic elevation (probably by folding) of Kulm rocks lying south of the Flözleeres trough.

Gabbroic intrusions and submarine volcanic activity occurred locally during Kulm deposition.

The deep-water trough was eliminated in the Upper Carboniferous (perhaps tectonically, perhaps by sedimentary infilling, or both), and a prolific northern source of sediment appeared, which provided material for the Coal Measures of the Ruhr and other areas in the Schiefergebirge.

Folding, uplift, and erosion occurred late in the Upper Carboniferous and/or during the Lower Permian. The Hesse highland was eroded to a low level and foundered beneath the Zechstein (Permian) sea. After further deposition in the Mesozoic, the whole region was uplifted and the sea expelled. Parts of Hesse, once a source of coarse boulders, now forms a graben which conserves the downfaulted remnants of the flat-lying strata which formerly covered the whole folded belt. The two structurally higher areas where the old folded tract is now exposed, the Schiefergebirge and Harz, were former sites of deep troughs. It is curious to note that the high and low positions occupied in the Paleozoic by the source land and depositional areas have since been fully reversed.

COMPARISON WITH THE CARPATHIAN FLYSCH

In discussing the Carpathian Flysch, Vašíček (1953) distinguishes pelitic "tegel," pelitic "schlier" with intercalated silty beds, and "flysch" with coarser graded beds. He accepts the theory of emplacement by turbidity currents for the coarser beds. But whereas Fiege imagined the finest beds to have been laid down at the greatest depth and farthest from the coast in a single extensive basin, Vašíček assumes that the fine-grained beds are commonly of shallow origin, or from more or less separate basins, the flysch representing the deposits of the deep part of the major troughs.

These two opinions are necessarily opposed in principle. Mapping of current directions and facies distribution and, where possible, closer stratigraphic correlation will have to be undertaken in each region of graded bedding to ascertain the paleogeographic relations. Perhaps some extensive schlier series originated beyond a flysch zone, with tegel and siliceous shales even farther away, as in the Kulm of Sauerland and Harz. In other cases tegel and schlier may prove to be deposits of more local basins removed from a supply of coarse grains, or bypassed by turbidity currents carrying sand. The tegel, according to Vašíček, is commonly rich in microfossils of the littoral and neritic zones, the schlier in those of the neritic zone. Such rocks are probably of near-shore origin.

The present authors can agree with Vašíček's contention that flysch-like deposits can originate in non-orogenic basins, for instance the present oceans. Vašíček, however, takes what we regard as an extreme view in claiming that flysch is not typical of geosynclines. A sufficient supply of coarse material and the required subaqueous slope will usually be combined only in a geosynclinal trough with adjacent geanticlinal source area. The great bulk of graded flysch series must therefore have been deposited in geosynclines. In the oceans turbidity currents may have been particularly active only during the low levels of the Pleistocene, but outside that period may have been infrequent, except in orogenically active areas. The deep oceans, therefore, should be used with caution when seeking a model of the basins in which fossil sediments were deposited. It is probable that many flysch series, even those with graded bedding, accumulated in basins of moderate size and in depths that did not exceed 1000 meters. The writers regard Vašíček's contention that flysch is restricted to the largest basins as an unwarranted generalization.

CONCLUSIONS

The following tentative conclusions can be drawn from our short reconnaissance of the Kulm and Flözleeres graywackes:

(1) The graded graywackes of the Flözleeres and Kulm of Sauerland and the Kulm of the Harz can best be explained by the mechanism of turbidity-current deposition. The interbedded conglomerates are due to slides.

(2) The general distribution of Kulm shallow-, slope- and deep-water deposits described by Fiege is confirmed by a number of measurements of turbidity-current directions and the distribution of sedimentary features.

(3) The area of the Hesse graben is shown to have been the source for the Kulm graywackes and conglomerates of the northeast Schiefergebirge and Oberharz, as suggested previously by Kühne and Paeckelmann, Stöcke, and others.

(4) The majority of flysch deposits, including Kulm and similar sediments, belong in the category of graded graywackes and were deposited at some depths. Locally, however, deposits of the same petrographic composition occur but lack grading. These appear to have formed closer inshore, and it can be tentatively suggested that the transportation took place on a steeper slope in the form of slides and turbidity currents and that the currents had not time while covering the short distance from their take-off to sort the materials according to grain size. The larger particles must first become concentrated in the nose of the current, ready to form the sole of the bed, before grading can be produced.

(5) In the cases examined, graded bedding becomes common at about 10 to 15 km from the shallow-water zone, and graywackes extend at least 40 km from the source. Assuming a slope of 5° over 10 km and 1° over the following 30 km, a depth of 1500 m is arrived at. This figure is suggested as a low estimate.

(6) Current-flow diections observed in the Flözleeres are opposed to the direction of source inferred by Fiege. Insufficient data were obtained by the writers to interpret the Flözleeres paleogeography with confidence.

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STRATIGRAPHIC RELATIONSHIPS OF THE METAMORPHIC ROCKS IN SOUTHEASTERN NEW YORK*

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ABSTRACT. The name *New York City group* is proposed for the sequence of metamorphic rocks consisting of the Fordham gneiss, Inwood marble, and Manhattan formation south of the Hudson Highlands in southeastern New York. The name *Manhattan schist* is considered inappropriate for the highest formation of the group because much of the formation is not schist.

The Fordham gneiss is a wholly conformable and integral part of the New York City group and cannot be considered equivalent to the Precambrian gneisses of the Hudson Highlands. Interlayering of Fordham, Inwood, and Manhattan type lithologies within formation belts reflects original alternation in type of sediments laid down. Such alternation in lithologic types is most pronounced close to the contacts between principal formation belts.

The so-called Lowerre quartzite, formerly considered to be the basal member of the New York City group, is not a valid formation. Examination of all reported occurrences of the Lowerre quartzite showed only two localities where quartzitic-looking rocks are present, and detailed studies showed these pseudoquartzites to be highly sheared phases of the Fordham gneiss and related intrusives.

Recognition of the Fordham gneiss as an integral part of the New York City group, and the non-existence of the "Lowerre quartzite," invalidate any correlation based upon presumed similarities in stratigraphic sequences between the New York City group and the Cambro-Ordovician series north of the Hudson Highlands. The New York City group may ultimately prove to be the correlative of the Cambro-Ordovician series north of the Hudson Highlands, but it seems likely that an acceptable correlation will have to be made by way of Connecticut, Massachusetts, and Vermont rather than directly across the Highlands.

INTRODUCTION

The problem of the relationship between the metamorphic terrane south of the Hudson Highlands in southeastern New York State and the variously metamorphosed Cambro-Ordovician series north of the Hudson Highlands has been studied for more than a century, but the best efforts of many geologists over the years have failed to present an unequivocal solution to the perplexing questions of geologic age and stratigraphic correlation.

Recent geologic mapping by the writer south of the Hudson Highlands in southern Putnam and northern Westchester Counties has yielded much information on the metamorphic rocks of that terrane and has served to clarify a number of relationships which must be understood before the larger regional problems can be solved. It is the purpose of this paper to set forth certain conclusions, substantiated by field and petrographic data, which are pertinent to an understanding of the regional picture.

Field work done by the writer during the summers of 1953, 1954 and 1955 was largely devoted to mapping the bedrock geology of the Lake Carmel, Brewster, Peach Lake, and Croton Falls 7½-minute quadrangles. Numerous brief investigations were made of the rocks in the surrounding areas. The writer was assisted in the field by Robert M. Sneider for three months in 1953 and 1954, by Robert M. Cassie for three months in 1955, and by John A. Graham for a total of about three months over the three-year period. The

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work, still in progress, is part of the basic geologic mapping program of the Geological Survey of the New York State Museum and Science Service.

METAMORPHIC ROCKS SOUTH OF THE HUDSON HIGHLANDS

General Relations

The terrane south of the Hudson Highlands, in Manhattan, Bronx, Westchester, and Putnam Counties, is underlain by a group of metamorphic rock formations consisting of the Fordham gneiss, the Inwood marble and the Manhattan "schist." The latter formation is comprised of abundant feldspathic gneisses and amphibolites in addition to true schists; hence, it is proposed that the more appropriate name *Manhattan formation* be used instead, as is done in this report.

The name *Manhattan group* has long been applied to these formations (Stevens, 1867, p. 120; Merrill, 1890, p. 390), although in late years the term has been little used. It is convenient to retain a group name for these formations, but the use of the same name for the group as for a formation within the group violates the 1933 code of classification and nomenclature of rock units of the American Commission on Stratigraphic Nomenclature (Art. 11, paragraph c). Although the term Manhattan group was proposed before Manhattan as a formation name, the latter use is more firmly established; therefore, it is proposed that the term Manhattan group be abandoned and the unit comprised of the Fordham gneiss, Inwood marble, and the Manhattan formation be renamed *New York City group.*

The Fordham gneiss, the lowermost unit of the New York City group, dominantly is a well banded hornblende-biotite-quartz-plagioclase gneiss having a granoblastic texture. Characteristically the individual gneiss layers are thin (<2 ft) and notably persistent along the strike. Interlayered with the gneiss are abundant amphibolite layers, marble, and mica schists. Both injection and replacement type granites are abundantly developed within the gneiss locally.

The Inwood marble is a white to gray, moderately silicated, medium-grained rock ranging in composition from calcite to nearly pure dolomite. The Inwood is indistinguishable lithologically from the marble layers occurring within the Fordham and Manhattan formations.

The Manhattan formation dominantly is a garnetiferous quartz-biotite-plagioclase gneiss characterized by abundant sillimanite and locally much muscovite. Its typical lepidoblastic texture is commonly so well developed locally that the term schist may be used appropriately, although as noted earlier the term is a misnomer when applied to the formation as a whole. Amphibolites and marble layers are abundantly present as interlayers, and in places the formation is intimately injected by granitic material. The sillimanitic phase of the Manhattan formation is its most definitive lithology and best serves to distinguish between the Fordham gneiss and Manhattan formation in places where there is otherwise a similarity of lithologic type.

The age of the New York City group has never been determined definitely. Mather (1843, p. 464) considered the group to be the metamorphic equivalent of the Paleozoic series north of the Hudson Highlands. Merrill

(1890, p. 391) also considered the group to be Paleozoic, but in 1898 he relegated the Fordham gneiss to the Precambrian and assigned the Inwood marble and Manhattan formation to the Ordovician. Berkey (1907, p. 377) considered the entire group to be Precambrian and later suggested it was equivalent to the Grenville series—a correlation principally based upon presumed similarities in rock types. Knopf and Jonas (1929, p. 71) correlated the Fordham gneiss with the Baltimore gneiss, which they considered to be Archean, and the younger formations of the New York City group with the supposedly Precambrian Glenarm series of Pennsylvania and Maryland.

The concept of the entire New York City group's being Precambrian persisted generally until the publication of Balk's classic Dutchess County studies (1936). Balk believed the Fordham gneiss to be the equivalent of the Precambrian gneisses of the Hudson Highlands, and that a complex system of semicircular faults divided the Fordham gneiss and younger New York City group formations into narrow parallel belts south of the Hudson Highlands. He correlated these younger formations with the Cambro-Ordovician series north of the Hudson Highlands (basal Poughquag quartzite, Wappinger limestone, Hudson River pelites).

Balk traced the increasing rank of metamorphism of the Poughquag-Wappinger-Hudson River pelite series eastward and attached great importance to the presumed identical stratigraphic order of that series with a higher-rank Lowerre quartzite-Inwood marble-Manhattan formation series lying southeast of the Hudson Highlands. He concluded that the two series are identical, that they are both Cambro-Ordovician, and that the Fordham gneiss is Precambrian and not an integral member of the original sedimentary series of the New York City group.

Relationship of the Fordham Gneiss

An understanding of the relationship of the Fordham gneiss to the other formations of the New York City group and to the Precambrian gneisses of the Hudson Highlands is a prerequisite to solving the stratigraphic correlation and structural problems of the extensive metamorphic terrane lying east and southeast of the Green Mountain anticlinorium axis.

The Fordham gneiss, the basal formation of the New York City group, is conformable to the Inwood marble and Manhattan formation. It may be traced almost continuously from New York City northeast and east to the vicinity of Danbury, Connecticut—a distance of about 60 miles—and exhibits throughout its extent a persistent stratigraphic position and wholly conformable relationship with the Inwood marble wherever exposed. Some workers have suggested that this apparent conformity is structural, but it appears unreasonable to expect that pseudoconformity produced by intense deformation would be so complete and persistent over such a great distance and that it would maintain so constant a stratigraphic position. Moderate to intense shearing, which has been noted at a few contacts of the Fordham gneiss with the Inwood limestone, is no more pronounced than one would expect in highly deformed rocks, particularly at the contact between rocks of widely different competencies. Such shear zones, in fact, are commonplace at many different

horizons within the group, and to accept such features as evidence of stresses which produced pseudoconformity of the Fordham gneiss seems unwarranted.

The presumed close similarity of the Fordham gneiss and the undoubted Precambrian gneisses of the Hudson Highlands led Balk (1936, p. 695, 772) and others to correlate the two. This appears to have been a major stumbling block in the structural interpretation of the area and has obfuscated certain critical points important to regional correlations.

It is not always possible, in the case of isolated outcrops, to distinguish between the Fordham gneiss and the Highlands gneisses, but wherever adequate outcrops permit careful observation of the general characteristics of the gneisses it is possible to distinguish the separate identities. Fluhr (1945; 1950, p. 182-184) was perhaps the first to call attention to the important differences in the gneisses.

Chief differences between the gneisses are: (1) the great persistence of thin inch-scale mineralogic layers in the Fordham in contrast to the ill-defined, discontinuous layers in the Highlands gneisses; (2) the abundance of relict sedimentary beds in the Fordham gneiss and the almost complete lack of them in the Highlands gneisses; (3) the presence of fracture and crush zones associated with well defined faulting and folding in the Fordham gneiss in contrast to the plastic flowage typical of deformation in the Highlands gneisses; (4) the general lack of clinopyroxenes in the Fordham gneiss in contrast to their abundance in the Highlands gneisses; and (5) the much less common and less advanced granitization of the Fordham gneiss compared to the Highlands gneisses.

The Fordham gneiss and the overlying Inwood marble and Manhattan formation are certainly valid formations in a regional sense and can be traced as recognizable units in proper stratigraphic sequence by reconnaissance mapping methods. Yet anyone who has tried to examine a number of individual outcrops within the formation belts, as for example in the New York City area, has probably been confused time and again to find "Manhattan-type" rocks in Fordham gneiss belts, and "Fordham-type" rocks in Manhattan formation belts, and "Inwood-type" rocks mixed in with both Fordham and Manhattan rocks.

Detailed outcrop mapping in the Croton Falls and Peach Lake quadrangles on a scale of two inches to the mile has shown that such interlayering of lithologic types is characteristic of the New York City group. Especially common are rocks of Inwood-type and Manhattan-type lithologies interlayered with the Fordham gneiss in the Fordham gneiss belts, and Inwood-type rocks interlayered with Manhattan formation in the Manhattan formation belts. This interlayering of lithologic types occurs on a wide range of scales, including interlayers no more than a few inches thick. Generally this type of interlayering is most abundant close to the contacts between the formational belts. Locally this alternation of rock types in the contact zones is so well developed that in large measure the actual contacts drawn on a map must be arbitrary.

In a few instances the occurrence of thick layers of Inwood-type marble and Manhattan-type formation within the Fordham gneiss belts possibly is the result of tight infolds of the higher formations into the Fordham gneiss.

In most cases, however, the lack of mirror repetition of rock types on opposite sides of the axis of any presumed infold militates against the interpretation of apparent interlayering being the result of severe folding.

But the intimate small-scale interlayering of distinct lithologic types is perhaps the most convincing evidence that the Fordham gneiss is an integral part of the metasedimentary New York City group.

One of the most interesting sections is that exposed along the Titicus River immediately below the Titicus Reservoir dam at Purdy's, New York. Here a continuous section 283 feet thick is exposed in New York City group rocks that are standing vertically. The sequence exposed is shown in simplified form in table 1.

TABLE 1

Sequence of Lithologic Types in Titicus Dam Section

Lithologic type	Thickness in feet
Manhattan	2.5
Inwood	1.5
Manhattan	69.5
Inwood	19.5
Fordham	54.0
Inwood	3.0
Fordham	54.0
Inwood	6.0
Fordham	1.0
Inwood	0.2
Fordham	21.0
Inwood	0.1
Fordham	2.0
Inwood	0.2
Fordham	49.0

The kind of interlayering described here can not be explained reasonably in terms of an unconformable relationship betwen the Fordham gneiss and the Inwood marble and Manhattan formation (see plates 1, 2, 3). It appears, rather, that the interlayering of the different lithologies reflects original alternation in types of sediments laid down. It reflects a shift from dirty sands through carbonates to argillaceous sediments, with many oscillations from one sedimentary type to the other during the time of deposition of the New York City group. As might be expected, the various combinations of sedimentary types differ from one locality to another and are most pronounced close to the contacts between the principal formation belts.

It is concluded that the Fordham gneiss is not correlative with the Precambrian gneisses of the Hudson Highlands but is an integral part of the metasedimentary group.

Status of the Lowerre Quartzite

The name *Lowerre quartzite* was introduced by Merrill (1898, p. 26) and was applied to a supposed thin-bedded quartzite overlying the Fordham

PLATE 1

Part of Titicus dam section showing conformable bands of rock with Manhattan-type (M), Inwood-type (I), and Fordham-type (F) lithologies.

gneiss near the old Lowerre railroad station in Yonkers. Merrill considered the so-called Lowerre quartzite to be the same as the basal quartzite of the Cambro-Ordovician series north of the Hudson Highlands, and in his New York City folio (1902, p. 4) he called the quartzite "Poughquag" because he believed it to be equivalent to the Poughquag quartzite of Dutchess County, which carries Lower Cambrian fossils and rests with sharp unconformity upon the Precambrian gneisses of the Hudson Highlands.

In addition to the presumed occurrence near Lowerre station, Merrill (1898, p. 26) cited occurrences in Hastings-on-Hudson at the Hastings marble quarry, and along the shore of the Hudson River at Sparta Landing, 1/3 mile south of Sing Sing prison in Ossining. Merrill's geologic maps included in the New York City folio show 9 other occurrences of the quartzite in the Harlem quadrangle.

Berkey (1907, p. 366-367) considered the Lowerre quartzite to be " . . . only an upper quartzitic facies of the basal or Fordham gneiss formation. In comparison with the great formations of the area it scarcely can claim a separate classification, but it is a distinguishable bed, the uppermost member, although not separate in any fundamental sense." In addition to the Sparta location, Berkey cited occurrences a mile south of Eastview along the Putnam division of the New York Central Railroad, in the small creek just south of

PLATE 2

Fordham gneiss (F) with conformable interlayer of Inwood-type marble (I) in Titicus dam section.

Crugers Station, and on the Harlem division of the New York Central Railroad about a mile north of Chappaqua.

Berkey and Rice (1921, p. 132 ff.) in a later report apparently considered the Lowerre quartzite to be a valid formation meriting unequivocal status in discussions of regional correlation, although mention was made of its erratic distribution. They considered the Lowerre to be Precambrian and probably Grenville.

Balk (1936, p. 689) correlated the supposed Lowerre quartzite with the Lower Cambrian Poughquag quartzite of the Cambro-Ordovician series north of the Hudson Highlands, which was a correlation based in large part upon the presumed similarity in stratigraphic sequence of the formations in the two series involved. The general excellence of Balk's paper did much to establish the Lowerre quartzite in the literature.

The absence of the so-called Lowerre quartzite in the quadrangles mapped by the writer led him to a more detailed investigation of the validity of the formation. The lack of any adequate field or petrographic descriptions of the unit in the literature, the newly established stratigraphic position of the Fordham gneiss, passing comment by other geologists familiar with the regional geology (T. W. Fluhr, 1950, p. 186; personal communication, 1954; K. E. Lowe, personal communication, 1954; Fluhr and Bird, 1939, p. 68),

PLATE 3

Thin interlayers of Inwood-type marble (I) in Fordham gneiss in Titicus dam section.

and Berkey's (1907, p. 367) early consideration of the Lowerre as merely an upper quartzitic facies of the Fordham gneiss added to a growing suspicion that the so-called Lowerre quartzite is not a valid formation and should not play a positive part in any regional correlations.

The writer has visited every locality mentioned in the literature as an occurrence of the so-called Lowerre quartzite. A few of the occurrences cited are no longer accessible, as they have been completely covered by pavement or buildings since first mentioned in the literature. At most places, however, there is every indication that the outcrops today are essentially the same ones present at the time they were cited in earlier reports. At only two localities, the Hastings marble quarry and along the Hudson River at Sparta Landing, did the writer find anything which even resembled a quartzite, and at these places field and petrographic studies clearly showed that the quartzite-appearing beds are in reality strongly sheared phases of the Fordham gneiss and related pegmatites.

Table 2 summarizes the nature and present status of reported occurrences of the so-called Lowerre quartzite. A somewhat detailed description of the occurrences at the Hastings marble quarry and at Sparta Landing follows.

TABLE 2

Reported Occurrences of the "Lowerre Quartzite"

Location	Citation	Description
Lowerre Station, Yonkers.	Merrill, 1898, p. 26	Abundant outcrops of Fordham gneiss along Lawrence St. Paving covers presumed stratigraphic position of the "Lowerre."
Hastings marble quarry, Hastings-on-Hudson.	"	Slabby, sheared phase of the Fordham gneiss.
Sparta Landing, Ossining.	"	Slabby, sheared phase of Fordham gneiss and related pegmatites.
Leake and Watts Orphan Asylum, Yonkers.	Merrill, 1902, Geol. map of Harlem quadrangle	Fordham gneiss and Manhattan formation in abundant outcrops. Inwood marble and stratigraphic position of the "Lowerre" are covered by pavement of Hawthorne Ave.
Yonkers, south of McLean Ave. between Tibbets Rd. and Sedgewick Ave.	"	Outcrops of Fordham gneiss and Manhattan formation; no "Lowerre."
Bronx, E. 204 st., between Villa Ave. and Grand Concourse.	"	Covered
Bronx, Morris Heights Station.	"	Covered
Bronx, Grand Concourse between Minerva Pl. and 199 st.	"	Covered
Bronx, Jerome Ave. at E. Burnside Ave.	"	Fordham gneiss exposed in new building excavation (4-55).
Bronxville, along Bronx River midway between Tuckahoe Rd. and Palmer Rd.	"	Fordham gneiss outcrops along river bank.
Yonkers, along N.Y.C.R.R., Putnam Div., south of intersection of Halladay Rd. and Palmer Rd.	"	Fordham gneiss outcrops in R.R. cuts.
Three-fifths mile north of Scarsdale Station, along Bronx River.	"	Fordham gneiss outcrops on hill west of river.
One mile south of Eastview, along N.Y.C.R.R., Putnam Div.	Berkey, 1907, p. 366	Abundant Fordham gneiss outcrops.
Crugers Station, in small creek along R.R. tracks.	"	Fordham gneiss outcrop in stream bank and along tracks to the north.
One mile north of Chappaqua on Harlem Div. of N.Y.C.R.R.	"	No outcrops along R.R. tracks, Fordham gneiss outcrops abundant along Saw Mill River Parkway to the west.

Hastings marble quarry.—The abandoned Hastings marble quarry is located in Hastings-on-Hudson, New York, along the east side of the Old Croton Aqueduct about 2000 feet north of the intersection of Pinecrest Drive and Warburton Avenue. The quarry was opened in the Inwood marble, which is here generally a light gray to white rock well foliated and somewhat friable. The marble strikes N 20°-40°E and is steeply overturned to the northwest. High up on the hanging wall, near the center of the old working face, is the contact between the Inwood marble and a slabby, light brown highly sheared rock considered by Merrill (1898, p. 26) and others to be the Lowerre quartzite. The contact of this pseudoquartzite with the marble formation is wholly conformable, but the contact zone appears to be a transition between the relatively pure marble in the quarry and the Fordham gneiss behind the hanging wall of the quarry. (The steeply dipping section is here slightly overturned to the northwest.) The nature of this transition is shown in the following partial section:

	Rock Unit	Lithology	Apparent thickness
Inwood Marble	Marble	Light gray to white, foliated, low in silicate impurities	>100′
	Diopside rock	Greenish	2′
	Siliceous marble	Light gray, quartz >50%, some lime-silicates	4′
Fordham gneiss	Granitic gneiss	Deeply weathered, schistose, crumbly or "rotten"	15′
	Granite pegmatite	Coarse, crumbly	6′
	Pseudoquartzite	Light brown, slabby, highly sheared with strong lineation	35′±
	Feldspathic gneiss	Well banded: alternating light and dark layers	>100′

The contact zone between the Fordham gneiss and the Inwood marble is a locus of intense shearing, as is commonly the case where two rocks of greatly different competencies are intensely folded together. The shearing is more evident in the more brittle rocks of the Fordham gneiss, but the silication and silicification of the Inwood marble close to the contact probably was localized by the shearing. The band of granitic gneiss 15 feet thick between the bottom of the Inwood marble and the pseudoquartzite places the latter well within the Fordham gneiss.

Petrographic examination of samples taken at random from the pseudoquartzite show it to be a strongly sheared, sub-mylonitized granitic gneiss, granite, or pegmatite. It is concluded that the so-called Lowerre quartzite here is not a sedimentary formation but rather a strongly sheared unit of the Fordham gneiss.

Sparta Landing.—This locality is along the Hudson River in the village of Ossining, about 2000 feet south of Sing Sing prison. A slabby, light brown quartzitic-appearing rock crops out along the New York Central Railroad at the base of the small hill immediately north of Sparta brook. Similar quartz-

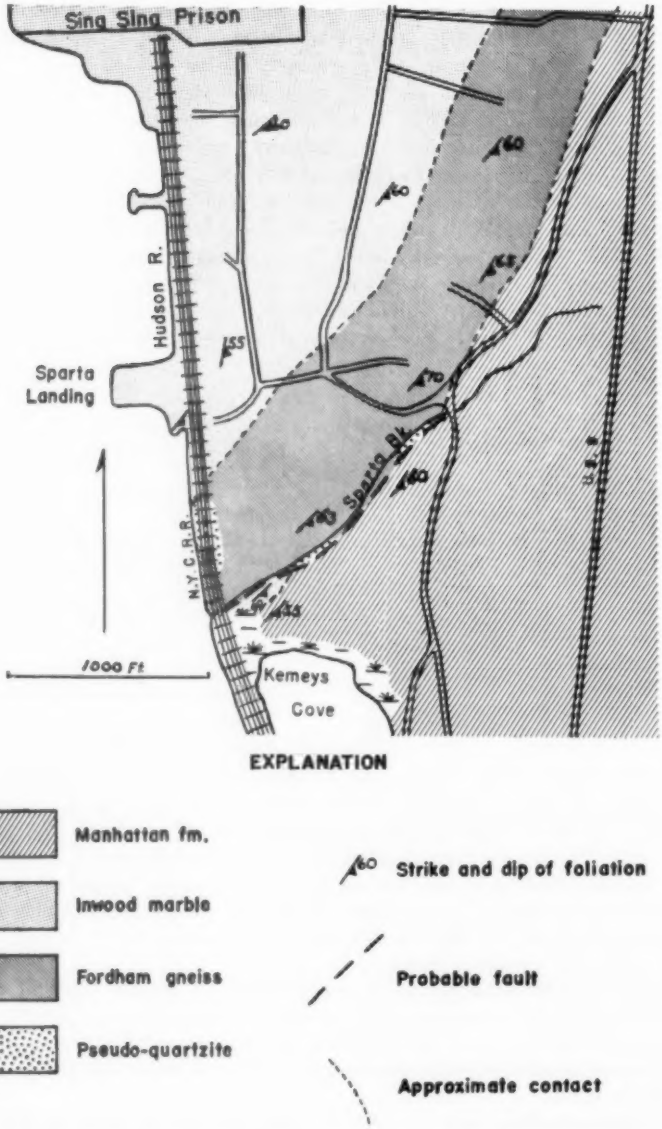

Fig. 1. Reconnaissance map showing geology in the vicinity of Sparta Landing, Ossining. North of Sparta brook the Inwood marble is believed to be squeezed out between the Fordham gneiss and Manhattan formation.

itic-appearing rocks occur along the southeast bank and in the bed of Sparta brook about 600 and 1250 feet above the mouth of the brook. The rock exposed along the railroad is strongly crumpled so that a precise determination of its thickness is not possible, but there is probably an apparent thickness of 100 feet or more. The exposed thickness of quartzitic-appearing rock along Sparta brook is 25-35 feet.

The small hill north of Sparta brook is underlain by a northeast-plunging anticline of Fordham gneiss which is abundantly injected by a younger microcline granite and related coarse pegmatites. North of the hill the bedrock is Inwood marble, which crops out at the landing at many places on the high ground east of the railroad. The hill southeast of Sparta brook is underlain by the Manhattan formation. Along the base of the western spur of this hill, on the north side near the brook, is a small outcrop of Inwood marble. The straightness of the brook along its lower reaches and the abundant evidence of shearing of the rocks along it strongly suggest a minor fault along which most of the Inwood marble between the Fordham gneiss and the Manhattan formation has been sheared out. The field relations are shown in figure 1.

Although casual field inspection of the quartzitic-appearing rocks might lead one to consider them to be true quartzites, closer inspection shows them to be strongly sheared phases of the Fordham gneiss and associated granite sheets. In several places massive granite could be followed through progressively intensified shearing into slabby pseudoquartzite. Examination of thin sections of the pseudoquartzite revealed it to have the mineral composition of a granite.

CONCLUSION

Recognition of the Fordham gneiss as an integral part of the metasedimentary New York City group and the non-existence of the Lowerre quartzite invalidate any correlation which is based upon presumed similarities in stratigraphic sequence between the New York City group and the Cambro-Ordovician series north of the Hudson Highlands.

This conclusion does not solve the difficult problem of the age of the New York City group, but it points up the fact that what is probably the most popular correlation accepted today for these rocks cannot be considered valid on the basis of the reasoning used.

The writer doubts that any satisfactory correlation of the New York City group will be made directly across the Hudson Highlands, although the New York City group may ultimately prove to be the correlative of the Cambro-Ordovician series north of the Hudson Highlands. It seems more likely that an acceptable correlation will have to be made by way of Connecticut, Massachusetts, and Vermont. The best correlations that can be made in that direction at the present time point to a lower Paleozoic age for the New York City group, but the route is circuitous and the evidence inconclusive.

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DEVONIAN REEFS FROM BELGIUM: RELATION BETWEEN GEOSYNCLINAL SUBSIDENCE AND HINTERLAND EROSION

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ABSTRACT. In the Upper Devonian of southern Belgium, the Frasnian shows rhythmic alternation of calcareous and shaly members, with three horizons of localized bioherm development. A detailed stratigraphic description by Lecompte permits an analysis of the crustal movements. The sedimentation rhythm is primarily controlled by tilting movements, in which a subsiding sedimentation basin is balanced by a rising hinterland. The marine sedimentation of the Frasnian was situated seawards of the hinge line of the tilting movement. This is in contrast to the paralic coal belts, which were situated landwards of the hinge line.

INTRODUCTION

Bioherms in the Upper Devonian of southern Belgium are a classical example of fossil reefs, repeatedly described and widely visited (Maillieux, 1922, 1933; Dumon, 1954). A very detailed study of Lecompte (1954) offers a means of analyzing the relationship in this area between changes of sedimentation and the formation of bioherms, which are together controlled by crustal movements during the Upper Devonian.

The reefs are found in three different levels of the Frasnian stage, which, with the younger Famennian, forms the Upper Devonian. Rhythms of various scales may be detected in the Devonian geosynclinal sediments. First, there is the larger rhythm, involving the whole Devonian, swinging from detrital-clastic in the Lower Devonian through the calcareous Middle Devonian back to the detrital-clastic Famennian. Second, there is a variation within the individual stages of the calcareous middle part of the system, between more purely calcareous series and series with terrigenous material, either marls or shales. Third, is a small-scale rhythm, found in many of the transition periods from calcareous to terrigenous sedimentation, in which individual beds with a higher lime content alternate with beds with more detrital material.

In the larger-scale rhythm, the Frasnian forms the transition between the mainly calcareous stages of the underlying Couvinian and Givetian and the purely detrital-clastic sedimentation of the overlying Famennian. It is formed by a series some 400 meters thick, exhibiting three main rhythms, each formed by a succession of shales, calcareous shales, and limestones, and superimposed upon the large-scale rhythm involving all of the Devonian. Apparently the reversal from calcareous sedimentation to detrital-clastic sedimentation in the larger-scale rhythm of the Devonian is very gradual, enveloping the entire Frasnian stage. During that stage, the facies repeatedly swayed from detrital-clastic to calcareous and back again, and continuous detrital-clastic sedimentation developed only in the Famennian. This gradual replacement of the calcareous facies by the detrital-clastic facies contrasts with the rather sudden transition from detrital-clastic to calcareous facies at the base of the Middle Devonian. That transition takes place in the calcareous shales of the "Schiste calcareuse de Bure," whose thickness of some tens of meters is slight com-

pared to the 3000 m of the Lower Devonian and the 1000 m of the Middle Devonian.

STRATIGRAPHY

Below the Famennian, which is developed as shales and sandstones, the Upper Frasnian is formed by the "Assise de Matagne," F 3. It is also developed in detrital-clastic facies, consisting of some 80 m of thinly bedded shales. No calcareous nodular shales are found, and no sessile fauna of any importance, such as corals or brachiopods, is recorded. A flimsy pelecypod, *Buchiola palmata*, and rare goniatites form the main fossils.

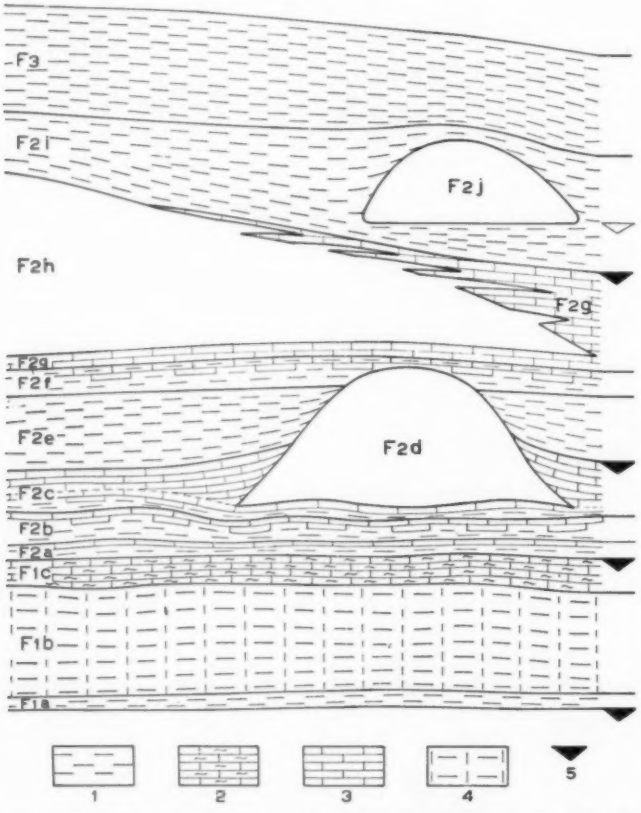

Fig. 1. Schematic stratigraphic section of the Frasnian near Frasnes. 1:2500.

Bioherms F 2 d, F 2 h and F 2 j white. In other sections F 2 d and F 2 j bioherms are much thicker, the total thickness of the bioherms of the three periods reaching mostly about 200 meters.

1—shale; 2—marly limestone; 3—well bedded limestone; 4—coarse-bedded limestone; 5—beginning of a period of subsidence (open triangle: subsidence of minor importance).

The main portion of the Frasnian is formed by the F 2, the "Assise de Frasnes." It comprises seven normal members, the F 2 a, F 2 b, F 2 c, F 2 f, F 2 g, and F 2 i, and three reef facies members F 2 d, F 2 h, and F 2 j, which locally develop in the normal members F 2 c, F 2 g, and F 2 i respectively. A summary of the normally stratified series is given first.

The F 2 i, the "Schistes à *Reticularia pachyrhyncha*," still is mainly detrital-clastic; it consists of thinly bedded shales and shales with calcareous nodules, with an impoverished fauna. In its middle part, this member shows somewhat stronger calcareous sedimentation, where thin limestone beds with brachiopods and corals alternate with the nodular shales. It is on these beds that the reefs F 2 j locally develop. There is no other relation, such as interfingering, between the massive reef limestone and the surrounding shales. The latter show distinct foreset bedding around the reef, which must have formed a submarine mount. The normal thickness of the F 2 i shales is about 60 m, rising to more than 100 m over an actual reef.

The F 2 g is the youngest mainly calcareous member, being built up by well bedded limestones and marly limestones, containing brachiopods and branching corals. This series, which thins out to less than 20 m thickness away from the reefs, attains a thickness of some 70 m near the reefs F 2 h. These, reaching more than 100 m themselves, stick out of the surrounding bedded strata of the F 2 g, and their upper part is covered by the shales of the F 2 i. The F 2 g itself contains much calcareo-detrital material from the actual reefs on their flanks, and is seen to interfinger with the reefs.

The F 2 f, grayish nodular and calcareous shales with echinoids, trilobites, brachiopods and pelecypods, is again mainly detrital-clastic. With its thickness of 10 m to 20 m it forms the transition between the still more clastic facies of the underlying F 2 e and the calcareous F 2 g. A small-scale rhythm is, however, superimposed here upon the main rhythm of the Frasnian, because the top beds of the F 2 e are formed by a thin horizon of nodular shales with branching corals, indicating a temporary period of shallower water.

The greenish, finely laminated shales of the F 2 e, developed in a thickness varying between 50 m and 100 m, do show some intercalations of slightly calcareous nodular shales, but they contain only an impoverished fauna of goniatites, gastropods and solitary corals. In the basal beds of this member, over a thickness not more than a couple of meters, the transition is found from the underlying calcareous F 2 c to the detrital-clastic F 2 e and F 2 f. The transition still contains a corallian fauna, accompanied by brachiopods, but stromatoporoids are absent. It must be noted that the transition from calcareous sedimentation to detrital-clastic sedimentation—from F 2 c to F 2 e—is much more sudden than the following reversal from detrital-clastic to calcareous—from F 2 e over F 2 f to F 2 g.

The F 2 c is formed by calcareous and nodular shales with a rich fauna of corals, brachiopods, and sponges. It attains some 30 m thickness outside the reefs F 2 d. Near the reefs it becomes much more calcareous, both by addition of calcareo-detrital material from the reefs and by development of the same corallian and stromatoporoid fauna that actually builds up the reef. Its thickness there rises to more than 50 m.

The F 2 b is again mainly detrital-clastic, consisting of 30 m to 40 m of greenish, finely laminated shales. There is, however, a gradual upwards augmentation of calcareous matter in its upper part, which first passes into nodular shales and then into calcareous shales, the latter with a rich fauna of brachiopods, pelecypods and solitary corals. Here again, the transition from the detrital-clastic facies towards the calcareous facies is gradual.

The F 2 a is also mainly detrital-clastic, being formed by some 10 m of nodular shales. In this thin member, a small-scale rhythm leads, however, to the formation of a couple of limestone beds at its top, containing a curious fauna of large-sized brachiopods. Here again, the transition from the calcareous intercalation of the "Zône des Monstres" towards the overlying detrital F 2 b shales is sudden.

The Lower Frasnian, the "Assise de Frommelennes," is divided into the F 1 c, F 1 b, and F 1 a. The F 1 c consists of 10 m to 20 m of limestones and marly limestones with brachiopods, pelecypods, and corals. Calcareous sedimentation is dominant, but some terrigenous influence is visible in the marls. Moreover, the stromatoporoids are absent. These are dominantly characteristic of the calcareous, thickly bedded F 1 b, where they develop beautiful biostromes. Evidences of emersion, such as the formation of "beach rock" and layers of single, disarticulated brachiopod shells, have been described, but no actual bioherms are found. The F 1 a, some 10 m thick, again marks a shaly intercalation between the stromatoporoid limestones of the underlying Givetien and those of the F 1 b.

The bioherms locally develop in the calcareous members F 2 c and F 2 g and on calcareous intercalations in the member F 2 i. The strata just below the bioherms become progressively richer in their fauna, whilst corals and sponges tend to predominate. The same fauna eventually constructs the base of the reef. At that time, there is no question of seamounts in the Devonian sea. If one would follow the French habit of speaking about the "hautfonds" on which the bioherms could form, these cannot have been more than the flattest of shoals. Outside the actual base of a bioherm, a similar accentuation of calcareous sessile fauna is found at the same time. The predominant parameter is time—the installation, at a given moment, of a shallow, well aerated sea, rich in nutrients and poor in detrital-clastic material. In this sea, bioherms did develop locally. After some time, detrital-clastic sedimentation again began, gradually enveloping the bioherms which now indeed did form submarine hills above the normally stratified surroundings. Eventually the bioherms too were drowned by the general subsidence which began at the onset of detrital-clastic sedimentation, and subsequent clastic sedimentation leveled off the existing differences in height. Consequently, during the next following period of calcareous sedimentation, the sea presented an entirely new topography. Bioherms did again develop locally, but generally at other localities. Although several hundreds of bioherms are known in the area, in only a few instances is a younger reef found on top of an older one.

The oldest reefs, the F 2 d, either begin as a corallian structure, which grades upwards into a bioherm dominated by stromatoporoids, or they may be formed mainly by the latter from the onset. Their structure is, however, not

so well known as that of the younger reefs in which many quarries have been opened.

The F 2 h reefs are similar, but do always begin as a corallian bioherm. Arriving at a certain altitude, stromatoporoids flourish, and both groups together build the upper structure. From here, detrital calcareous matter plays a large part, whilst the bioherm is found to interfinger with the surrounding normal strata. Wave action evidently has become of importance, degrading the organic structure. It follows that the lower corallian structure developed below a marked wave activity, which points to a bathymetric difference in corallian and stromatoporoid habitat, the latter developing in shallower water.

Similar indications have been found in the normally stratified members, where abundant stromatoporoid development is related to emersion or shallow water phenomena. Corals on the other hand, although present in the shallow water biotope, thrive also in deeper water, where stromatoporoids are absent.

The reefs F 2 j also begin with a corallian structure. In their middle part, some detrital calcareous matter is found, together with nests of brachiopods and lamellar ("not very healthy") colonies of stromatoporoids. Some wave action clearly was felt but much less than in the F 2 h reefs. Moreover, no detrital calcareous matter was transported outside the actual reef, nor did it interfinger with the surroundings. The F 2 j reefs are capped by a thin sequence of corallian or crinoid limestone, marking the deepening of the sea on top of the reef, which evidently could not construct itself fast enough to offset the general subsidence.

CRUSTAL MOVEMENTS

The rhythmic alternation of detrital-clastic and calcareous members in the Frasnian is related to differences in depth of the Devonian sea. The detrital-clastic sedimentation occurred in deeper water than the formation of the limestones and marls. In the latter, moreover, one can distinguish between beds with a predominantly corallian fauna and those with stromatoporoids plus corals. The latter have been formed in the shallowest environment. The changes from deeper water to shallower water and back again in Frasnian time were continuous at a given place but do show a difference of speed. The change from detrital-clastic to calcareous sedimentation is much more gradual than the rather sudden change from calcareous to detrital-clastic beds. The latter switchover normally takes place in only one of two meters of sediment. The bioherms are also affected by the transitions from shallower to deeper water, but they show a time lag. They are able to offset the subsidence of the sea floor for some time by individual constructive growth.

The change from calcareous to detrital-clastic sedimentation is thus related to a rather sudden relative subsidence of the sea floor. Subsidence continues during the deposition of the detrital-clastic member, but it gradually lessens. This allows sedimentation to catch up with subsidence, whereby the water depth diminishes, until shallow water calcareous sedimentation again takes over.

Further analysis of the crustal movements at the time of the transition from calcareous to detrital-clastic sedimentation leads to the following result. The sudden subsidence of the sea floor at that time cannot have been caused by a general subsidence of the crust, nor by an eustatic rise of the sea level. Both phenomena would have led to the same effect in sedimentation, viz. a deeper water environment only would have been formed. They cannot have led to the establishment of a regime of detrital-clastic sedimentation, because a general relative lowering of the crust does not provide the clastic material which ultimately had to be laid down in the sedimentation basin.

The Ardennes region was already thoroughly base-leveled by Frasnian time. Postorogenetic crustal movements related to the Caledonian orogeny had led during the Lower Devonian to an encroachment from the North of the Old Red Continent. These movements had, however, entirely abated already during the Middle Devonian. An eustatic rise of the sea level, or a general subsidence of the crust, would only have led to a deepening of the wide shelf seas and a drowning of the adjacent base-leveled continent. Nowhere would erosion in this case be able to produce the clastics, sedimented in the shaly members of the Frasnian.

The fact that every change-over from shallow water sedimentation to deeper water sedimentation is coupled to a change-over from calcareous to detrital-clastic sedimentation proves that the subsidence of the sea floor in the sedimentation basin is the result of tilting movements of the crust that produced, at the same time, a rise of the hinterland. Renewed erosion there could then furnish the clastic material, which was not at hand during the previous period of calcareous sedimentation.

DEVONIAN SEA AND CARBONIFEROUS COAL BELT

Consequently, crustal movements in the sedimentation basin of the Devonian sea during the Frasnian were comparable to those in many paralic Carboniferous coal belts. There, also, analysis of the interrelation between sedimentation level and sedimentation character in the rhythmic cyclothems leads to the postulating of tilting movements of the Earth's crust. These also caused a rise in the hinterland coupled with subsidence in the adjoining ocean (Rutten, 1952).

There is, however, a marked difference between the paralic coal belts and the marine sedimentation basin of the Frasnian in the position of the sedimentation area with respect to the hinge line of the tilting crustal movements. In the paralic coal belt the sedimentation area lies landwards of the hinge line. When the hinterland rises—a fact inferred from the coarser sediment—the sedimentation area rises too, thereby causing a regression. In the marine Frasnian, the sedimentation area lies seawards of the hinge line. A rising movement of the hinterland—as deduced from the advent of detrital-clastic material—is coupled with a subsidence of the sedimentation basin. This in turn leads to a deeper water environment and a temporal extinction of the shallow water fauna. First to go are the stromatoporoids, then the corals and sponges, next the brachiopods and crinoids, until ultimately the whole sessile fauna vanishes.

Perhaps the contrast between the mainly continental sedimentation in a paralic coal belt and the marine sedimentation in a basin like that of the Frasnian is caused by this difference in position in regard to the hinge line. Perhaps the landward position of the paralic sedimentation basin automatically leads to a somewhat higher mean topographical situation, as compared to a marine sedimentation basin, situated seawards of the hinge line. This is, at present, a rather hypothetical question, which can be answered only if and when more types of basins have been analyzed in as much detail as the coal belts and the Frasnian in southern Belgium.

At present it seems better to stress only that detailed analysis of rhythmic sedimentation in the Frasnian leads to the postulation of tilting crustal movements during the Devonian that are comparable to similar movements deduced from analysis of sedimentation in a coal belt. Even so, differences exist between the Devonian and the Carboniferous sequence. In the Carboniferous, two modes of crustal movement were found, one a general subsidence, the other a tilting movement. The general subsidence, which could not yet be detected in the Devonian, is actually the more sudden of the two.

Rhythmic cyclothem movements of the Carboniferous have been studied in paralic series, where slight crustal movements are readily reflected by a different facies. It seems probable that the marine Devonian sequence has not so faithfully recorded every crustal movement, though cyclothem rhythms might have led to variations in detail, such as the intercalation of the "Zône des Monstres" in the F 2 a. In that case, a number of cyclothem rhythms might be lumped together in each member of the Frasnian which might be comparable to the megacyclothems of the coal belts. The thickness of the individual members of the Frasnian, in general greater than most coal belt cyclothems, makes this the more probable. In the all-marine environment of the Frasnian the minor cyclothem rhythms would then not be so well expressed, whilst only major movements, probably comparable to those separating megacyclothems, made themselves felt.

To the smaller cyclothem rhythm and the larger Frasnian rhythm, another can be added, viz. that of the main Devonian rhythm. As cited in the introduction, the change-over from detrital-clastic to calcareous sedimentation is sudden in the main Devonian rhythm. In the Frasnian rhythms, in contrast, it is gradual, whilst here the change-over from calcareous to detrital-clastic is sudden.

These, at present, are theoretical considerations only. It is well, however, to retain, first that the Devonian rhythms are due to tilting crustal movements, second that they show differences with coal belt cyclothems, and third that they probably represent larger rhythms than cyclothem rhythm. It follows that crustal movements may be very varied, and that extrapolation from one special type is hazardous.

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THE ROLE OF CREEP AND RAINWASH ON THE RETREAT OF BADLAND SLOPES

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ABSTRACT. In badlands near Wall, South Dakota, two distinct types of topography have developed. The broadly rounded interfluves on the Chadron formation are interpreted as due largely to creep, whereas the steep, straight slopes formed on the overlying Brule formation are attributed to dominance of rainwash erosion.

Measurements of erosion depth along stake profiles on badland slopes reveal that erosion is rapid, ranging from 0.8 to 1.5 inches under 32 inches of rainfall during a 25½-month period. Erosion depth increases with slope angle on Brule slopes controlled by rainwash, but is a maximum on the convex divides of the highly permeable Chadron slopes. Pediments adjacent to the badland slopes were lowered 0.9 inch during the same period.

The topographic differences appear to resemble those existing between characteristic humid and arid topographic types, or between the ideal forms postulated by Davis and Penck in their respective cycles of slope retreat. It is suggested that the dominance of creep over rainwash erosion causes the development of broadly rounded divides and convex slopes, whereas predominant rainwash maintains steep, straight, parallel-retreating slopes heading in narrow divides. The contrasting forms observed may be viewed as end members of a continuous series.

INTRODUCTION

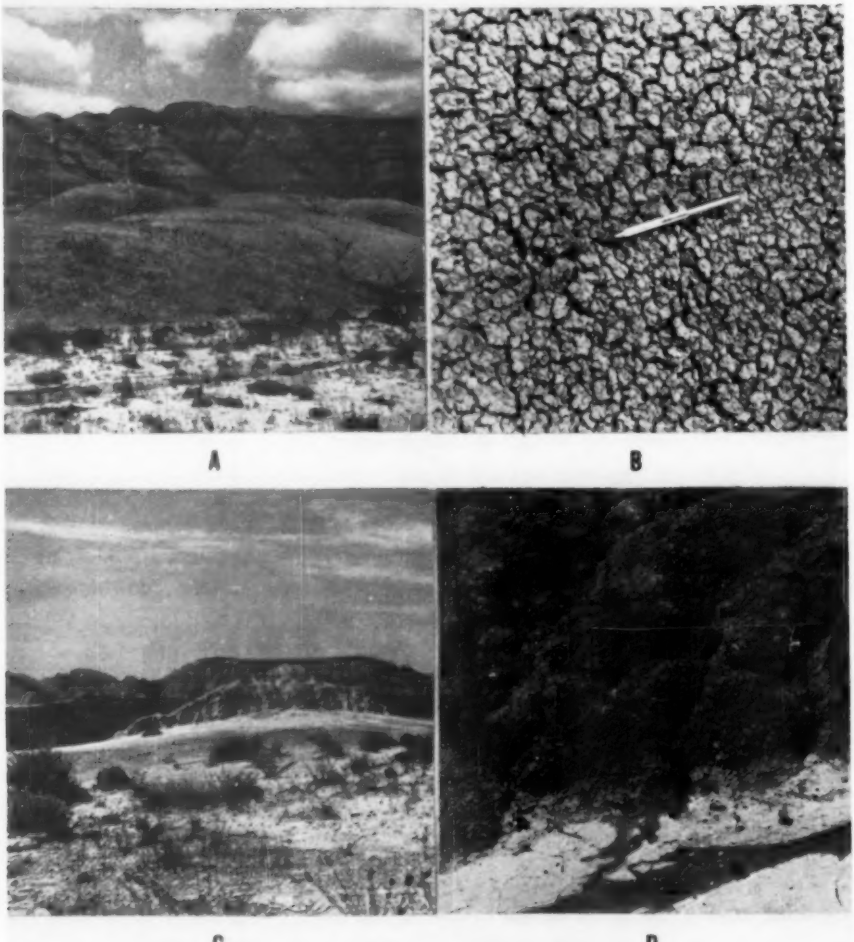
In the past the assignment of roles of creep and rainwash in the development of landforms has occasioned some controversy. The importance of creep in forming convex hilltops has been emphasized by Davis (1892), Gilbert (1909), Baulig (1940), Birot (1949, p. 23), and Penck (1953, p. 78), and the importance of rainwash stressed by Fenneman (1908) and Lawson (1932). Both concepts have been summarized by Baulig (1940), Cotton (1952), and Van Burkalow (1945).

Creep has been defined by Bryan (1922, p. 87) as "the slow movement of soil and rock waste down the slope from which these materials have been derived by weathering. Creep is due primarily to gravity but is facilitated by the presence of water, alternate wetting and drying, freezing and thawing, growth and decay of roots, and the work of burrowing animals." Gilbert (1909) generalized the mechanism of creep in the statement that "whatever disturbs the arrangement of particles permitting any motion among them promotes flow or creep."

Rainwash has been defined by Bryan (1922, p. 89) as "the water from rain, after it has fallen on the surface of the ground and before it has been concentrated into definite streams." Rainwash is the "unconcentrated wash" of Fenneman (1908).

During investigations of slopes and drainage systems of badlands in South Dakota, Arizona, and New Jersey (Schumm, 1956), examples were found which may illustrate the contrasting action of creep and rainwash. In Badlands National Monument, South Dakota, two distinct types of topography have developed. Steep, sharp-crested slopes, on which the mean of a sample of maximum slope angles measured 44°, have formed on the Brule formation, whereas the topography developed on the underlying Chadron formation, where exposed by the retreat of the overlying Brule, is composed of broadly rounded interfluves with a mean maximum slope angle of only 33° (pl. 1-A).

PLATE 1



Topography and slope surfaces at Badlands National Monument, South Dakota.

A. Typical topography of the Brule and Chadron formations. Contact between the two formations, just above the center of the photograph, separates the steep slopes developed on the Brule formation from the gentle, rolling topography developed on the Chadron formation.

B. Closeup of loose mosaic of aggregates forming the surface of slopes developed on the Chadron formation.

C. Small residual of the Brule formation showing straight steep slopes despite short slope length. Note pediment surface rising to make a sharp junction with the residual slope.

D. Water poured onto the surface of the Chadron formation follows subsurface channels to the base of the slope where it reappears on the pediment surface, sapping back the base of the slope.

The surface appearance of the slopes developed on each of the two formations is markedly different. In the dry state the Brule surface is hard; rill channels testify to copious runoff, although desiccation cracks are numerous. The dry Chadron surface, on the other hand, is composed of a loose mass of clay aggregates (pl. 1-B) which may be scooped up by hand.

The rapidity of breakdown of the Chadron clay and the looseness of its surface might suggest rapid erosion, and yet the harder, seemingly more resistant Brule has retreated far back from the margin of the Chadron (pl. 1-A). The Chadron thus appears to act as a more resistant rock. An explanation was apparent when a canteen was emptied on both types of slopes. Runoff occurred almost immediately on the Brule slope; whereas the water was completely absorbed by the loose Chadron aggregates. This suggested that the Brule slopes retreat more rapidly under the action of surface runoff, or rainwash, while the Chadron slopes with high infiltration rates are modified more slowly by the action of creep. To check these conclusions a 5-gallon hand pump was used to spray water on the slopes. On slopes developed on the Brule formation runoff usually occurred almost immediately and in all cases before 1 gallon of water was sprayed on an area of about 6 square feet. The water was quickly concentrated into rill channels and flowed away. In contrast 4½ gallons were sprayed onto about 6 square feet of Chadron slope before runoff began. By then the Chadron surface had become a sticky mud on which each aggregate had softened and slumped down until it came in contact with the next lower aggregate on the slope. From plate 1-B it is possible to visualize how the aggregates would move, sliding on the underlying fine-grained surface which apparently becomes sealed when wet.

EROSION MEASURED ON BADLAND SLOPES

Erosion was measured on badland slopes developed in a clay-sand fill at Perth Amboy, New Jersey (Schumm, 1956) by driving wooden stakes into the slopes along profile lines orthogonal to the contours of the slope. The stakes were oriented normal to the surface and driven flush with the slope surface. Erosion depth was measured by the length of stake exposed after each storm. Using similar techniques, stake profiles were established on some slopes in the South Dakota badlands in July 1953. The stakes were 18-inch segments of concrete reinforcing rods, ⅜ inch in diameter.

After the installation of the profiles the areas were revisited 15 months (October 1954) and 25½ months (August 1955) later. Between July 1953 and October 1954 precipitation was about 20 inches but only 12 inches during the second period. The total was thus about 32 inches of rainfall between July 1953 and August 1955 at Badlands National Monument Headquarters (J. H. Fraser, personal communication).

Stake profiles are plotted in figures 1 and 2 with the measured erosion for the two periods indicated above each stake. Profiles A and B on figure 1 are from small residuals developed on the Chadron formation. Profiles A and B of figure 2 are on a larger residual composed of both the Brule and Chadron formations. Stake 4 of profile A (fig. 2) is on the contact between the two

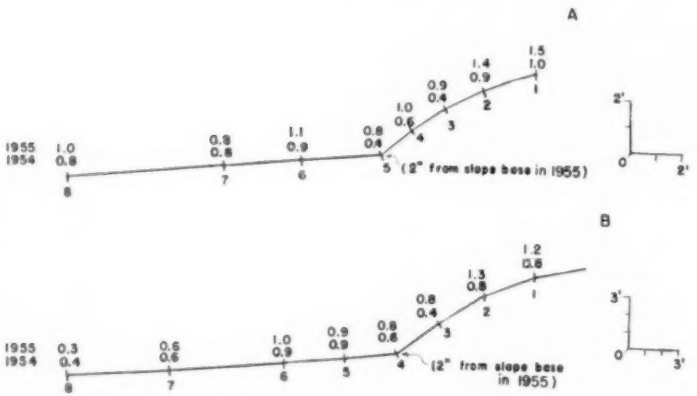


Fig. 1. Profiles showing slopes and pediments developed on two small badland residuals composed of the Chadron formation. Ticks on profiles show position of stakes: numbers above profiles indicate depth of erosion in inches measured in 1954 and 1955. The number assigned to each stake is listed below the profile. During the period of study the slopes of the two profiles have retreated 2 inches from the stakes driven at the junction of pediment and slope (stake 5, profile A; stake 4, profile B).

formations. The last vestige of Brule shale has been removed above profile B (fig. 2) which represents a slope composed entirely of the Chadron formation.

No stake profiles were established wholly on the relatively resistant Brule formation residuals with low infiltration rates, but profiles established on the fill at Perth Amboy illustrate erosion on slopes of similar appearance under the action of rainwash. At Perth Amboy the steep, straight slopes whose mean maximum angle is 49° intersect in narrow-crested interfluves very similar in appearance to those of the Brule. Among the many slope profiles measured at Perth Amboy that were straight from crest to base were several with convex summits. In these cases the parallel-retreating straight slopes had not entirely planed away remnants of the original upland surface. The break in slope profile between the upland and bordering straight slope was rounded by rainwash passing over the break in slope. The three Perth Amboy slope profiles selected for comparison with the Chadron profiles have convex summits and are illustrated in profiles A, B, and C of figure 3. The rainwash erosion on the relatively impermeable, convex Perth Amboy slopes may then be compared with the action of creep on the permeable, convex Chadron slopes.

Comparison of the profiles reveals differences in erosion rates between slopes here attributed predominantly to creep (profiles A and B, fig. 1; profile B, fig. 2) and those on which rainwash is assumed to be dominant (fig. 3 and possibly the combination slope fig. 2-A). On the Perth Amboy profiles (fig. 3) erosion at any point appears to be related to slope inclination. On the steep, straight portions of the slope, erosion is at a maximum and decreases on the convex summit areas. Compared to erosion on the summits, erosion on the straight portions of the slope is essentially uniform. Differences in erosion depth exist on the straight portions, but for the most part are due, as

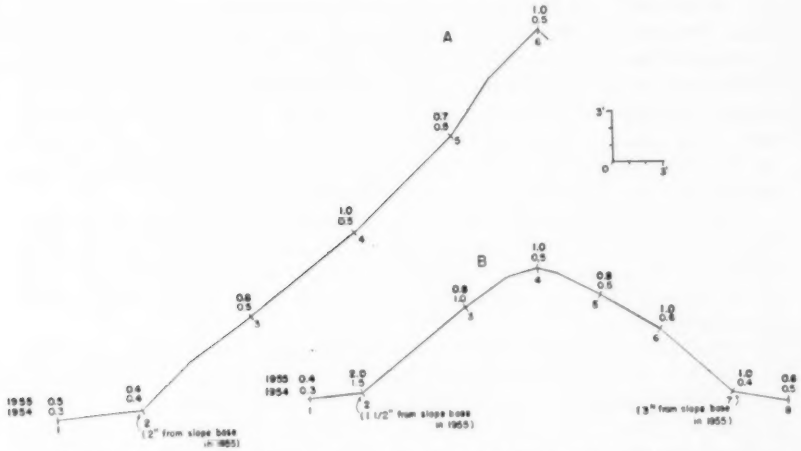


Fig. 2. A. Profile of slope composed of both the Brule and Chadron formations. Stake 4 is driven into the slope at the contact between the two formations. Stake numbers are given below each profile; erosion depth in inches above each profile.

B. Profile across same residual as in A above. Brule formation has been eroded off leaving slopes composed entirely of Chadron formation.

at stake 9, figure 3-B, to a slight concavity or in other cases to a convexity on the slope surface. A statistical comparison of erosion on 16 slopes during the 10-week period and of slope angles measured at the same points in 1949 and 1952 at Perth Amboy cast no doubt upon the hypothesis that the steep, straight portions are retreating parallel under the action of rainwash (Schumm, 1956).

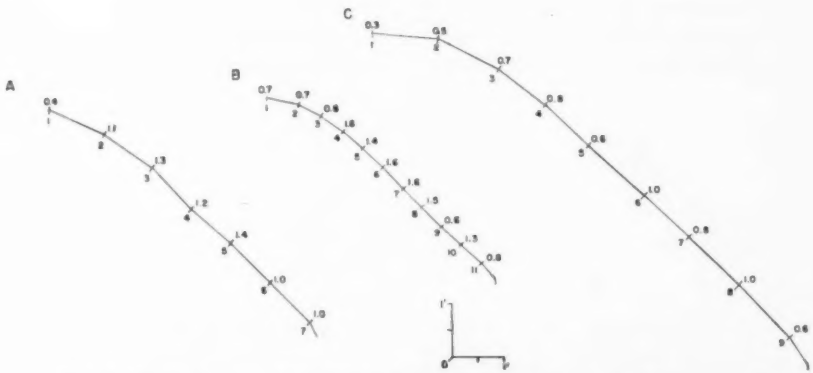


Fig. 3. Perth Amboy, New Jersey, badland slope profiles. Numbers above each profile give measured depth of erosion for a 10-week period in 1952. Ticks on profiles indicate position of stakes.

Erosion on the Chadron residuals of figure 1 shows an opposite relationship. Total erosion depths measured after 15 and 25½ months are greatest on slopes of least inclination. On profile B, figure 2, no systematic relationship is apparent between slope angle and erosion depth, yet a convexity is apparently developing. Note that stake 3 on profile B showed no erosion for the latter 10½-month period, but rather deposition of 0.2 inch. The lack of stakes on the lower steeper parts of this profile precludes the acquisition of data critical for the understanding of mass removal on the slope.

On profile A, figure 2, erosion was uniform over the length of the slope during the first period of erosion and is essentially uniform for the 25½-month period despite the fact that the mean slope below the contact between the Brule and Chadron formations (stake 4) is 39° and that above on the Brule formation is 44°.

The opposite relations as to erosion depths displayed by Chadron profiles in comparison with those of the Perth Amboy badlands (fig. 3) are considered most important to this discussion. Reversed relationships of slope steepness to erosion depth in these two areas can perhaps be reconciled by appealing to dominance of two different geomorphic processes: creep and rainwash.

Runoff or rainwash where observed on badland slopes occurs as surge or subdivided flow (Horton, 1945; Schumm, 1956). Minute obstacles check and deflect the runoff in its downslope course; the movement of the eroded material is therefore not continuous. The runoff neither constantly accelerates nor becomes overloaded toward the slope base on straight slopes in the cases observed. Erosion measured on straight badland slopes is essentially uniform over the length of that slope, although during any one storm the mean erosion depth for the slope as a whole may not be reached or may be exceeded on any part of the slope.

If the slope has a convex summit, due in this case to rounding of the junction between the valley-side slope and original upland slope by runoff, erosion will be less on the gentler slopes, and the steeper portion of the slope will encroach on this convexity until two straight slopes unite to form a sharp crest. In other words, when rainwash acts on a slope any particles A, B, C,

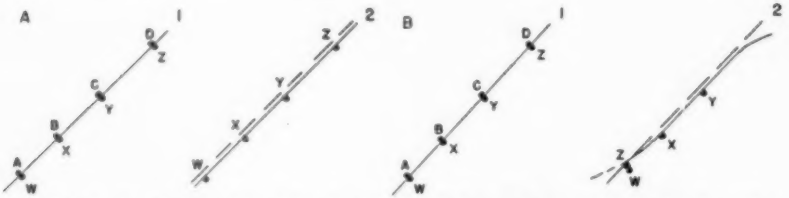


Fig. 4. A. Erosion of particles A, B, C, and D from a slope surface under the action of rainwash. The grains are not necessarily removed in sequence, but each layer of grains is removed before erosion of the next can begin.

B. Removal of particles A, B, C, and D from a slope surface by creep. The grains were removed in sequence, followed by particle Z, causing maximum erosion on the crest of the slope.

and D (fig. 4-A) spaced equally along the slope may be removed from the slope during any one storm or during several storms, but in any case before the underlying particles W, X, Y, and Z are removed. Rainwash is thus an eroding mechanism tending to remove a uniform thickness of material from a straight slope during a storm. It does this because as an eroding agent it attacks the steepest part of the slope with the greatest energy. Thus, when particle A (fig. 4-A) is removed from the slope, in effect, the base level for the particles above is lowered and they are quickly removed. If by chance particle D is removed first, a slight concavity results and erosion is less there until the lower particles on the slope have been removed. This type of action has been documented by repeated measurements along the stake profiles at Perth Amboy (Schumm, 1956).

Creep has been proposed as the dominant process active on the Chadron residuals. The individual aggregates move downslope by swelling from wetting, shrinking from drying and filling of the desiccation cracks from the uphill side, and possibly by rainbeat and sliding and slumping on the sealed, underlying, finer-grained surface. Unlike the action of rainwash, in which a sheet of material moves on the entire slope surface, a particle D (fig. 4-B) engaged in creep will not be removed from the slope at the same time or before a particle A but must traverse the entire slope length behind the particle downslope from it. Thus, the aggregates will be removed from the slope in the order A, B, C, D, Z. Since the movement of the creeping material should be greatest on the steepest slopes, it would seem that erosion would logically be greatest on these slopes. However, it may be that these slopes are principally slopes of transportation. The creeping material from upslope moves over these slopes, protecting the bedrock from rapid erosion.

Perhaps the above explains the partial burial of stake 3 on profile B of figure 2. Furthermore, a stake profile established on a badland slope in Sioux County, Nebraska also illustrates the partial burial of the stakes on steeper slopes (fig. 5). The slope illustrated in figure 5 is steep, bordered at its base by an ephemeral stream, and its surface is composed of highly permeable aggregates derived by weathering of the Chadron formation. After 1 year the upper two stakes of the profile were exposed by the action of creep. Of the

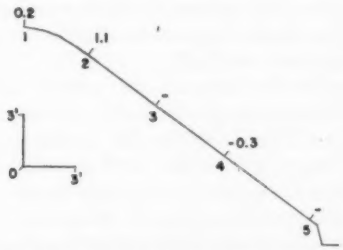


Fig. 5. Profile of badland slope developed in the Chadron formation (Sioux Co., Nebraska). Erosion is greatest at the upper two stakes, while the lower three were buried by downward creeping material from the slope crest. Erosion depth in inches is given above the profile.

lower three stakes on the steepest part of the slope, only stake 4 was found, buried 0.3 inch. Stakes 3 and 5 are assumed to have been buried as well.

If, as observed on slopes modified by creep, wasting is most rapid on the gentler slopes near the summit of a hill, the convexity will broaden and encroach on the less rapidly eroding side slopes, as illustrated by the lowering of the Chadron residuals. A lune-shaped wedge of material is thus removed from the slope as indicated by Lawson (1932) but not, as he thought, by rainwash. Perhaps, also, the supply of material from the flatter summit slopes and the ease of movement on these slopes is increased by the greater amount of rain falling per unit area as compared with the steeper side slopes

EROSION ON MINIATURE PEDIMENTS

Stake profiles were extended across the miniature pediments bordering the badland residuals in figure 1-A and B. In each case the pediment was lowered, presumably by rainwash. Similar reduction of miniature pediment surfaces has been described by Bradley (1940) and Smith (in press).

In this area Smith recognized that while some pediments were being dissected and lowered others were being raised by aggradation. He concludes that the sheet wash is not a major erosional agent but is chiefly one of transportation. He expects slow lowering of the pediment, however, when it is graded to a "slowly falling base level stream."

In profile the pediments are slightly concave-up. Both pediments are graded to a small ephemeral stream. McGee (1897) and Davis (1938) agree that under such conditions the pediment probably will be degrading and probably concave in longitudinal profile.

The hypothesis favored here is that although modified by rainwash the pediment is not formed by it. Both Paige (1912) and Davis (1938) considered that formation of the pediment is by retreat of a steeper slope leaving the pediment behind as a surface of transportation to the nearest drainage channel or interior basin. In the South Dakota badlands the miniature pediments lie at the bases of retreating badland slopes. Slope retreat is accelerated by undercutting of the upper slope which differs on the two lithologic units. Greater runoff on the Brule shale slopes and "the spreading of sheetwash at rill channel mouths undercuts the escarpments" (Smith, in press). On pediments formed at the base of Chadron residuals the slope retreat is aided by basal sapping of the slope by the appearance of subsurface flow at the junction of slope and pediment surface (pl. 1-D).

During 25½ months the bases of the residual slopes retreated on an average of 2.0 inches from the stake originally placed at the junction of pediment and badland slope (figs. 1 and 2). The pediments were lowered on an average of 0.7 inch during the first 15-month period but only 0.2 inch more during the second period for a mean of 0.9 inch for the 25½ months. During the second period deposition occurred at the lower stake of profile B, figure 1. It may be that the greater total rainfall of the first period, including one storm which totaled 3.46 inches during 24 hours, produced much greater runoff which lowered the pediments by sweeping depositional material from the pediment surface.

EROSION IN ARID AND HUMID CLIMATES

The differences observed between the two types of badland topography in Badlands National Monument have been interpreted as due to the different processes acting on the slopes: creep and rainwash. Because the topographic differences appear to resemble those existing between characteristic landforms of humid and arid climates, the explanation of the differences in the badlands may be projected to include differences in humid and arid topographic forms. Figure 6 shows the writer's concept of slope development on the highly permeable Chadron formation contrasted with that on the impermeable Brule formation. The differences are perhaps also illustrative of the distinctions between the classical slope retreat concepts of W. M. Davis and W. Penck. Under the action of rainwash (fig. 6-A) the initial slopes retreat at an essentially constant angle, leaving a pediment at their base adjusted to the angle required to allow removal of the debris brought to the slope base, thereby

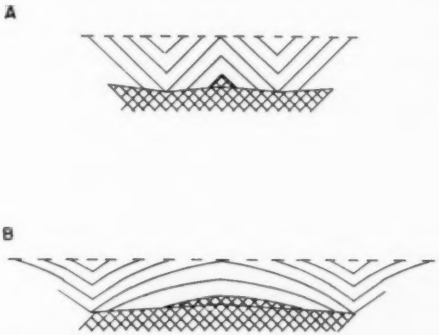


Fig. 6. A. The cycle of slope development on the Brule formation.
B. The cycle of slope development on the Chadron formation.

perpetuating the maintenance of the straight slope. The slopes remain constant until only a small residual remains surrounded by a pediment surface (pl. 1-C).

Slope angles measured on Brule residuals of all sizes are essentially constant in comparison to those measured on Chadron residuals, substantiating parallel slope retreat in this type of landform evolution (Schumm, 1956). Some reduction of slope angle with erosion of the Brule shale slopes may occur, however, due to the alternation of resistant and less resistant layers. The slopes capped by or composed of several resistant layers are somewhat steeper than those where the resistant layers have been removed by a shortening of the slope. Smith (in press) has shown a slight but significant lowering of slope angles on slopes ranging in length from 45 to 3 feet. He attributes this decrease in angle to a lessening of the undercutting of the slope base as it becomes shorter, causing a declining slope retreat even on the Brule formation residuals. This is not substantiated by the basal slope retreat measured on the profiles of figures 1 and 2, which is about 2 inches for all the slopes.

In the development of slopes controlled by creep (fig. 6-B), rounding of the upper slopes occurs early in the cycle, and the radius of curvature increases through much of the later slope development. In this example (fig. 6-B), however, illustrating the development of the Chadron residuals, a pediment also forms at the base of the steeper lower slope. In this area the pediment is apparently formed by basal sapping of the slope by the appearance of subsurface flow at the slope base (pl. 1-D). If this basal sapping did not occur transport would be slowed at the base of the steep slope, and deposition there would probably aid the formation of a concave lower slope analogous to the concave basal slopes of humid climates.

In an effort to obtain some information on the progress of the erosion cycle on the Brule and Chadron formations, a series of slope profiles was measured on remnant buttes exemplifying the results of erosion on slopes of each type.

In figure 7 three profiles were measured across the divide on a residual composed of the Brule formation. The mean slopes of each profile are not markedly different; a small decrease in mean angle is the result of the removal of a nodular layer which causes the convex appearance of the right slope of profile 1. Areas of more impermeable Brule maintain even steeper slopes until late in the cycle (pl. 1-C).

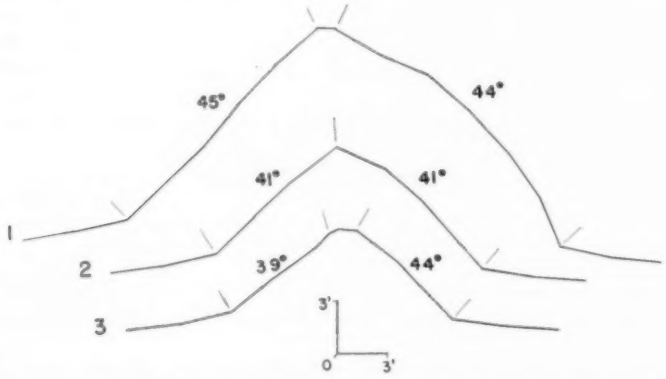


Fig. 7. A series of slope profiles measured on a residual of the Brule formation. Mean slope angles change only slightly with shortening of the slope on this relatively impermeable material.

Figure 8 shows the profiles of figure 2 with two additional profiles added to show the decline of the mean slope angle after the removal of the overlying Brule shale. Profile 1 (fig. 8) is a slope composed of both the Brule and Chadron formations. The dashed line indicates the contact between the overlying Brule and the stratigraphically lower Chadron. With the removal of the Brule from the upper part of the slopes, profiles 2, 3, and 4 show a progressive decline in mean slope angle.

Figure 9 shows a series of profiles across a residual of highly permeable Chadron clay. The texture of the topography in this area is relatively coarse

for badlands, as would be expected in areas of permeable soils. The profiles show the marked reduction in slope angle as the erosion of the residual progresses.

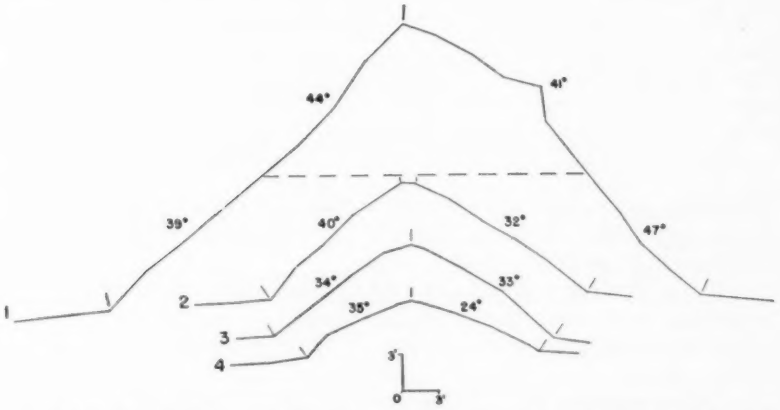


Fig. 8. Series of slope profiles on the residual where the stake profiles of figure 2 are located. The lower three (Chadron) profiles have moderate infiltration rates in comparison to the low rate of figure 7 profiles and the high rate of figure 9 profiles. Dashed line (profile 1) indicates contact between the overlying Brule and the Chadron formation.

The differences between erosion on the Brule and Chadron formations may be more clearly illustrated by comparing the rate of lowering of the residual divides with that of the reduction in width of the residual. In the three profiles of figure 7 (Brule) between profile 1 and 3 the width of the residual was reduced by 50 percent and the height by 56 percent. On the Chadron profiles of figure 8 between profile 2 and 4 the width was reduced only 56 percent while the height was lowered by 72 percent. On figure 9 (Chadron) between profiles 1 and 3 the width was reduced by 63 percent and the height by 81 percent. Therefore, in spite of variations in slope angles caused by lithologic units of differing resistance, the height of the Brule formation residual was lowered only 6 percent more than the width, indicating a small decrease in slope angle, whereas the height of the Chadron residuals is decreased by 28 percent and 18 percent over the width, causing a marked decrease in slope angle. These series of slope profiles seem to substantiate the hypothetical cycle illustrated in figure 6 and indicate that in this badland area the topographic differences are most reasonably explained by the difference in infiltration rates of the slope surfaces.

A further distinction between landforms fashioned by creep and those eroded by rainwash is suggested by a comparison of the rates of erosion on the Perth Amboy, New Jersey badlands and Chadron slopes in Badlands National Monument. The contrasts in mean slope angles and texture may be considered analogous to differences between any two areas of high relative relief fashioned by the two processes. Also, the comparison is justified on the

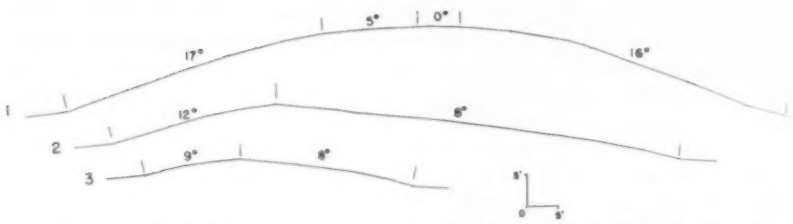


Fig. 9. Series of slope profiles on a badland residual located in a highly permeable area of the Chadron formation.

assumption that the clay-sand fill at Perth Amboy is probably about as easily eroded as any relatively impermeable material, while the Chadron is a rapidly eroding area in which permeability is high.

On the Chadron slopes 32 inches of rain fell, while a mean depth of 1.1 inches of material was removed. At Perth Amboy the mean depth of erosion was 0.9 inch in a period in which about 15 inches of rain fell (U. S. Weather Bureau, 1952).

Comparing directly, the Perth Amboy slopes were reduced an average of 0.2 inch less than the Chadron but with 17 inches less rainfall. With equivalent rainfall over the two areas (32 inches) the Perth Amboy slopes would, at the same ratio, be eroded to an average depth of about 2.0 inches as compared to the 1.1 inches on the Chadron. In other words, slope reduction would proceed at a rate about twice as rapidly by rainwash as by creep.

Influencing the reliability of this estimate are many factors of which the most important is probably the precipitation intensities involved. It is, however, reasonable to suppose that in areas otherwise comparable, rainwash erosion will proceed at rates nearly double that by creep. The hills of Chadron formation lying in front of the retreating Brule escarpment in the South Dakota badlands seem to support, at least qualitatively, the above estimate.

This concept may be of importance in attempting to extend the relationships between the geomorphic and hydrologic characteristics of drainage basins in semiarid regions to more humid areas and also as a partial explanation of the high rates of erosion in semiarid regions.

CONCLUSION

In humid regions rapid rock disintegration produces thick soil horizons protected by good vegetation cover. As Davis (1930) has suggested, creep is probably dominant there. Abundant evidence (Sharpe, 1938) has been cited of its importance. In arid regions with poor vegetation cover and thin soils, rainwash is probably the dominant process. Undoubtedly, both occur in each climatic region, but in humid areas rainwash erosion probably occurs only in exceptional storms; whereas in arid regions creep is important probably only after several consecutive rains have saturated the soil. It may be that rainwash erosion and parallel slope retreat are dominant in the early stages of the erosion cycle in humid regions, when relief is great and slopes are steep

(Birot, 1949, p. 23), but that with the development of thick soil cover and grading of the slopes the action of creep and declining slope retreat become the characteristic process and form.

It is possible that the areas in which creep dominates over rainwash, or vice versa, are end members of a continuous series ranging through all proportions of both processes depending on prevailing vegetation, soils, and climate. Holmes (1955) has suggested that the classic examples of the arid and humid geomorphic cycles, as contrasted in the Penck and Davis concepts of slope retreat, are end members of such a series. This concept appears to be supported by the badland studies.

If creep is responsible for convex interfluves in the mature and later stages of erosion (after any initial surface of low relief is completely consumed), there is no need to resort to Penck's explanation that the recent diastrophic history of a region controls the slope profiles. In Penck's system the straight slopes of the Brule formation residuals would indicate uniform uplift (gleichformige Entwicklung) whereas the convex slopes of the Chadron residuals would indicate increasing rates of uplift (aufsteigende Entwicklung). Both diastrophic histories cannot be true within the same small area. Penck claims that such local exceptions do not invalidate his concepts. Perhaps this is true, but by the same token the diastrophic control is merely an unnecessary complication whose introduction is unjustified in many areas.

ACKNOWLEDGMENTS

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THE SYSTEM Mg_2SiO_4—Mg_2GeO_4

A. E. RINGWOOD

ABSTRACT. Mg_2SiO_4 and Mg_2GeO_4 form a continuous series of solid solutions. The melting behavior is of the Roozeboom Type I. Similarity of melting point and latent heat of melting are indicative of the close thermochemical and structural relationship between the two end members. Attempts to prepare the spinel polymorph of Mg_2GeO_4 described by Goldschmidt were unsuccessful. It seems likely that the spinel modification is monotropic and is formed only under specialized and restricted conditions.

INTRODUCTION

In the course of a study of thermodynamic relationships between the olivine and spinel structures an investigation was made of phase equilibria in the system Mg_2SiO_4—Mg_2GeO_4. It was hoped that this system might provide significant data on the olivine-spinel transition, since Goldschmidt (1931) had previously reported the existence of both spinel and olivine type polymorphs of Mg_2GeO_4. In this respect, however, the results were disappointing and alternative methods of approach were found necessary. These will be reported upon in subsequent publications.

EXPERIMENTAL

The pure components MgO, GeO_2, and SiO_2 were thoroughly ground together under benzene in the desired proportions and formed into 1 gram tablets in a small press. Tablets of the composition S_{100}, $S_{80}G_{20}$, $S_{60}G_{40}$, $S_{40}G_{60}$, $S_{20}G_{80}$, G_{100} were prepared ($S = Mg_2SiO_4$, $G = Mg_2GeO_4$). The subscripts refer to weight percent. The tablets were sintered at 1350°C for 4 hours, then reground, and the process repeated twice.

Small quantities of samples (½ mg) were placed on an iridium strip heated by low voltage A.C. controlled by a variable transformer and voltage stabilizer. The temperatures of beginning of melting were determined by a Leeds and Northrup Optical Pyrometer focused on the strip immediately adjacent to the sample. The temperatures obtained on the pyrometer were calibrated against the melting temperature of Mg_2SiO_4 (1890°C). The precision of temperature measurement is low—possible errors of the order of ±30°C. are to be expected.

Refractive indices (alpha and gamma) of samples were determined with the aid of immersion liquids.

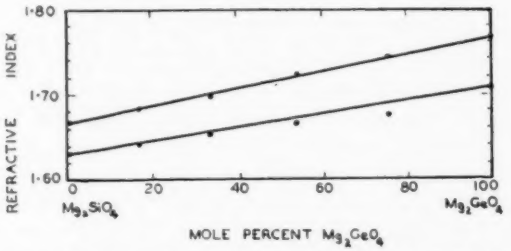

Fig. 1. Alpha and gamma refractive indices of Mg_2SiO_4-Mg_2GeO_4 solid solutions.

RESULTS

When the α and γ refractive indices are plotted on a mole percentage basis the variation is found to be essentially linear (fig. 1). The slight deviation from linearity displayed by some samples is probably due to inhomogeneity caused by differential volatilization of GeO_2. These results indicate that Mg_2SiO_4 and Mg_2GeO_4 form a complete series of solid solutions.

The temperatures of beginning of melting of the solid solutions are recorded in table 2. These temperatures are very similar. The melting temperature of Mg_2GeO_4 is only slightly below that of Mg_2SiO_4. There is no minimum (or maximum) at an intermediate composition (within the limits of error of temperature measurement). Consequently the phase relationships must be of the Roozeboom I type. The melting interval is observed to be very small; hence the solidus and liquidus must be very close.

The close crystal chemical relationship between germanates and silicates was first demonstrated by Goldschmidt (1931). This has been substantiated by the results of numerous other workers, e.g. Goldsmith (1950), Ingerson, Morey, and Tuttle (1948). This similarity also extends to thermochemical properties. Melting temperatures of silicates and germanates are usually very close, and in the case of the Mg_2SiO_4—Mg_2GeO_4 system, latent heats of melting must also be close. The widespread resemblance of germanates and corresponding silicates was explained by Goldschmidt (1931) as being due to the similarity of ionic radii (Si^{4+} .42Å; Ge^{4+} .53Å).

TABLE 1

Refractive Indices of Solid Solutions in the System Mg_2SiO_4—Mg_2GeO_4

Composition	α RI ($\pm$.003)	γ
Mg_2SiO_4 (S)	1.635	1.670
$S_{80}G_{20}$	1.645	1.682
$S_{60}G_{40}$	1.653	1.697
$S_{40}G_{60}$	1.666	1.725
$S_{20}G_{80}$	1.678	1.744
Mg_2GeO_4 (G)	1.708	1.765

TABLE 2

Temperature of Beginning of Melting of Mg_2SiO_4-Mg_2GeO_4 Solid Solutions ($\pm$30°C)

Composition	°C
Mg_2SiO_4 (S)	1890
$S_{80}G_{20}$	1865
$S_{60}G_{40}$	1860
$S_{40}G_{60}$	1850
$S_{20}G_{80}$	1845
Mg_2GeO_4 (G)	1855

THE SPINEL-OLIVINE TRANSITION

Goldschmidt (1931) reported that when Mg_2GeO_4 was precipitated from aqueous solution by Müller's method, pressed, and annealed at 600-800°C, it was found to possess the spinel structure. The mention was in a small footnote, and no further details were provided. When Goldschmidt prepared Mg_2GeO_4 at high temperatures, it possessed the olivine structure. The occurrence of polymorphism in Mg_2GeO_4 aroused some interest in view of the possibility that Mg_2SiO_4 might undergo a similar transition in the Earth's mantle.

Jander and Stamm (1932) synthesized Mg_2GeO_4 by heating a pastille of intimately mixed oxides at 1050°C for several days. On measuring the temperature variation of electrical conductivity of these samples they observed a sharp discontinuity (with rising temperature) at 1065°C and concluded that Mg_2GeO_4 undergoes a sluggish transition from a spinel to an olivine modification at this temperature. They could not reproduce this effect with falling temperature.

Attempts by the author to invert Mg_2GeO_4 from the olivine to the spinel structure were unsuccessful. Samples of Mg_2GeO_4, both pure, and containing 2 percent of lithium fluoride were held at 1000°C, 900°C, and 800°C for 12 days. No change was observed. Other mineralizers were tried—sodium fluoride, sodium tungstate, lithium carbonate, but without success.

When larger amounts (10-50%) of LiF were added to the charges, considerable amounts of euhedral octahedral crystals were sometimes observed. These were either isotropic or else they displayed very faint birefringence and polysynthetic twinning. Refractive indices were variable ranging from 1.75 (+) to 1.70 (−) according to the amount of LiF. It is possible that these crystals may consist largely of cubic Mg_2GeO_4 containing Li_4GeO_4 in solid solution.

Subsequent to this work, the author's attention was drawn to two recently published accounts of attempts to prepare cubic Mg_2GeO_4. Urey (1952) records that attempts to prepare the cubic modification of Mg_2GeO_4 in his laboratory and in Zachariasen's laboratory at Chicago had failed. Romeijn (1953) found that at 1200°C the solid solubility of Mg_2GeO_4 in Co_2GeO_4 (which has a spinel structure) was only 10-15 percent. With greater concentration of Mg_2GeO_4, olivines were formed. Furthermore, when he synthesized Mg_2GeO_4 by solid state reaction at 1200°C and 850°C, he found that it possessed the olivine structure. Romeijn concluded that "most probably Jander and Stamm are wrong on the dimorphism of magnesium germanate."

Because of these divergent views, further work was carried out by the author. Mg_2GeO_4 was precipitated from aqueous solution using Müller's method. (This was formerly a standard analytical method for the estimation of germanium.) Two-tenths of a gram of GeO_2 was fused to a glass, then crushed and dissolved in 100 cc of water. This was rendered slightly acidic with H_2SO_4, 20 cc of 2N ammonium sulphate solution and 20 cc of N magnesium sulphate solution were added, followed by ammonium hydroxide until

an excess of 20 cc was present. The liquid was boiled, and the precipitate filtered, washed with NH_4OH, and dried. The precipitate was then formed into a tablet in a press.

Samples of the precipitate were held at 110°C, 300°C, 500° C for ten days and others were held at 650°C, 750°C for five days. Powder photographs were then taken of the samples using Cu radiation. Powder patterns of the first three samples revealed about 4 or 5 diffuse lines together with a number of faint sharp lines. The last two samples gave patterns with many lines corresponding to olivine type Mg_2GeO_4.

The diffuse reflections at 500°C were found to correspond approximately to the strongest lines of the normal Mg_2GeO_4 olivine pattern, whilst the weak but sharp set of lines were found to belong to MgO. This is understandable since it is now known that when Mg_2GeO_4 is precipitated from aqueous solution some excess MgO is carried down (Kraus and Johnson, 1953). Because of this Müller's method has been superseded as a means of quantitative analysis.

The evidence cited above indicates that the olivine structure is probably thermodynamically stable for Mg_2GeO_4 at temperatures above 600°C. Accordingly the author would agree with Romeijn that Jander and Stamm were probably wrong about the transition of Mg_2GeO_4 at 1065°C.

The conflicting findings regarding the existence of a spinel modification are difficult to explain. The repeated failure of Urey and the author to prepare the spinel over a wide range of temperatures in the manner mentioned by Goldschmidt suggests that it was originally formed under rather specialized conditions and is most probably a low temperature, thermodynamically unstable (monotropic) form. In view of the paucity of information regarding the means of preparation supplied by Goldschmidt, this seems at present the most likely explanation.[1]

Goldschmidt quotes Dr. Hauptmann as finding a lattice constant of 8.3 Å for cubic Mg_2GeO_4. This value seems unlikely. The lattice constant of Ni_2GeO_4 (spinel) found by Romeijn and confirmed by the author (unpublished work) is 8.221 Å. Since the ionic radius of Mg^{++} (.66 Å) is smaller than that of Ni^{++}(.69Å), it is unlikely that the lattice constant of spinel type Mg_2GeO_4 will exceed 8.22 Å. Romeijn shows that when Mg^{++} and Ni^{++} substitute for one another in equivalent positions in spinels the lattice parameter remains practically constant. Unpublished work by the author on Co_2GeO_4—Mg_2GeO_4 solid solutions is in agreement with this finding. When the lattice parameters

[1] Note added in press: This conclusion may require modification in the light of recent work by Roy and Roy (1954) which became available to the author after the above paper was written. These authors repeatedly synthesized Mg_2GeO_4 spinel under hydrothermal conditions below 1000°C, and observed that it transformed very sluggishly to olivine at temperatures above 1005°C. Apparently, the transition was not observed in the reverse direction. The conflict in the experimental data is evidently due to the sluggishness of the transition and the consequent metastability of either the olivine or spinel below 1000°C. Until the transition is proved to be reversible the question must remain open.

of these solid solutions are extrapolated, the lattice parameter of cubic Mg_2GeO_4 is found to be 8.23 Å in good agreement with that of Ni_2GeO_4 and significantly less than Hauptmann's value.

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